Flexural extension of the upper continental crust in collisional foredeeps

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ABSTRACT

Normal faults on the outer slopes of trenches and collisional foredeeps reveal that high-amplitude lithospheric flexure can result in inelastic extensional deformation of the convex side of a flexed plate. This process, which we call "flexural extension," differs fundamentally from rifting in that the lower lithosphere contracts while the upper lithosphere extends. In the Taconic foreland of New York, a >100-km-wide zone of brittle failure propagated ahead of the convergent plate boundary, rupturing the upper crust to an estimated depth of 15–20 km. Dip-slip displacement on normal faults in the Taconic and Arkoma foredeeps produced water depths like those in the closest modern analogue, the Timor Trough. Structural evidence does not support common illustrations of flexural normal faults as planar-irrotational structures which simply die out at shallow crustal depths. Instead, the surface geology shows that flexural normal faulting must be rotational with respect to the enveloping surface of the flexed plate. This toppled domino geometry implies the presence at depth of a detachment or zone of distributed ductile simple shear where fault displacement and block rotation are accommodated.

INTRODUCTION

Although deep-sea trenches mark convergent-plate boundaries, the outer slopes of trenches are zones of extension, not contraction. Normal faults are clearly evident in bathymetric and seismic reflection profiles (Ludwig and others, 1966), and associated earthquakes consistently yield extensional first motions (Stauder, 1968). This apparent contradiction—extension in the direction of plate convergence—was explained by Ludwig and others (1966) as a response of the downgoing plate to bending. Most present workers accept this basic rationale, although Karig and others (1976) noted that there are two bends in a typical subduction zone, a gentler one seaward of the trench, characterized by normal faulting, and a sharper one beneath the forearc, where the plate dives into the mantle. An influential early illustration by Isacks and others (1968) showed normal faults in outer trench slopes as crustal-scale analogues of downward-terminating tension cracks that form on the convex side of a finite-thickness folded layer (Figs. 1A and 1B). Later depictions of subducting plates (for example, Karig and Sharman, 1975) typically show normal faults that either die out downward at a few kilometers depth (Fig. 1C), or that continue below the section to end in an unspecified way at an unspecified depth.

Like oceanic trenches, the outer slopes of some collisional foredeeps on continental lithosphere are also cut by normal faults that result from

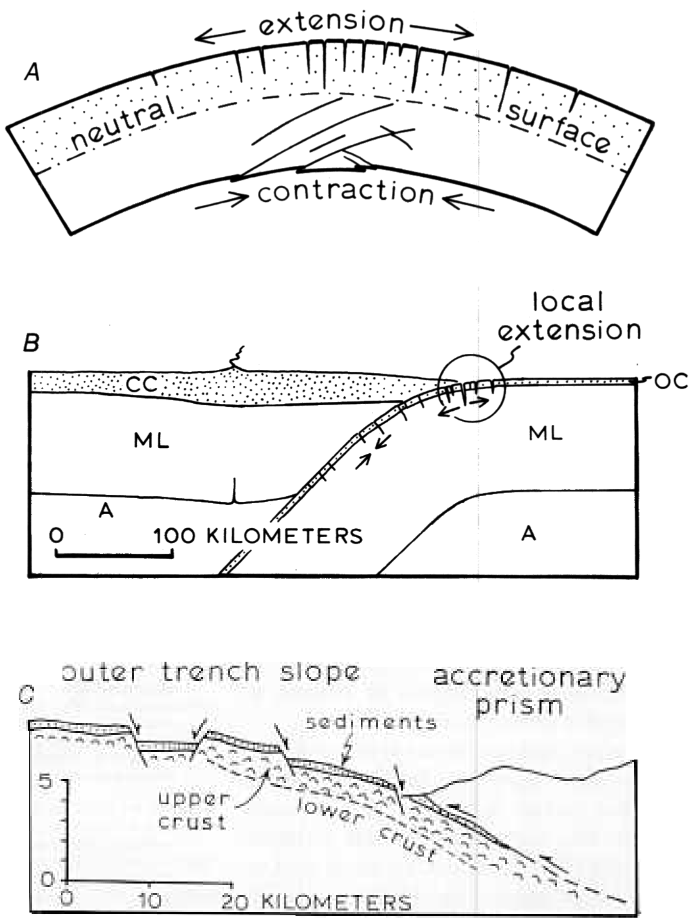

Figure 1. (A) Extensional deformation in the convex outer arc of this experimentally deformed material is accommodated by the opening of dilatant cracks that die out downward toward a neutral surface (from Ramsay, 1967, p. 401; no scale shown in original diagram). (B) Schematic representation of crustal-scale tension gashes formed as a result of lithospheric flexure at a subduction zone, adapted from an illustration by Isacks and others (1968). Arrows show opening and closing of dilatant cracks as the plate bends, then unbends. A is asthenosphere; ML is mantle lithosphere; OC is oceanic crust; CC is continental crust. (C) Typical representation of a normal faulted outer trench slope showing horst and graben topography and downward disappearance of normal faults (from Karig and Sharman, 1975).

Additional material for this article (appendix and two figures) may be obtained free of charge by requesting Supplementary Data 9127 from the GSA Documents Secretary.

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Flood-plain aggradation appears to be caused by large summer or fall floods, and flood-plain preservation seems to be aided by decreased flood frequency, particularly in winter.

Decreases in flood frequency, duration of high daily mean discharge, and changes in flood-plain alluviation are explained by low-frequency climatic fluctuations. Flood frequency decreased after 1942 in response to a decrease in the frequency of dissipating tropical cyclones and frontal systems. Intensity of precipitation decreased after about 1942, although seasonal amounts of precipitation show no temporal trend. The close relation between changes in frequency of floods, storms, and flood-plain alluviation strongly suggests that climatic fluctuations have a large effect on aggradation and degradation of the flood plain and channel and on sediment load in the Paria River.

Suspended-sediment load and flow volume of the Paria River have a distinct seasonality that is related to regional climatic conditions. Storm type, size, and frequency all affect sediment load and flow volume. About 90% of the annual sediment load is transported in summer and fall seasons (July 4–September 3 and September 4–November 9, respectively) when monsoonal thunderstorms, cutoff lows, or tropical cyclones are the dominant storm types. Only about 7% of the annual load is carried in the winter (November 10–April 17), when most precipitation is produced by frontal systems. Although little sediment is transported in winter, winter flow accounts for almost 50% of the annual flow volume. An insignificant amount of sediment is transported in spring (April 18–July 3).

Variations from year to year in both sediment load and flow volume have been greatest in the fall, when changes in regional climatic conditions also have been greatest. El Niño–Southern Oscillation (ENSO) conditions, which tend to produce large storms from dissipating tropical cyclones, have contributed to the greater variability in the fall.

Sediment storage in tributary flood plains such as the Paria and Little Colorado Rivers could have accounted for much of the decrease in sediment load in the Colorado River in the early 1940s. Flood-plain storage in the Paria River basin probably is controlled largely by variations in regional climatic conditions.

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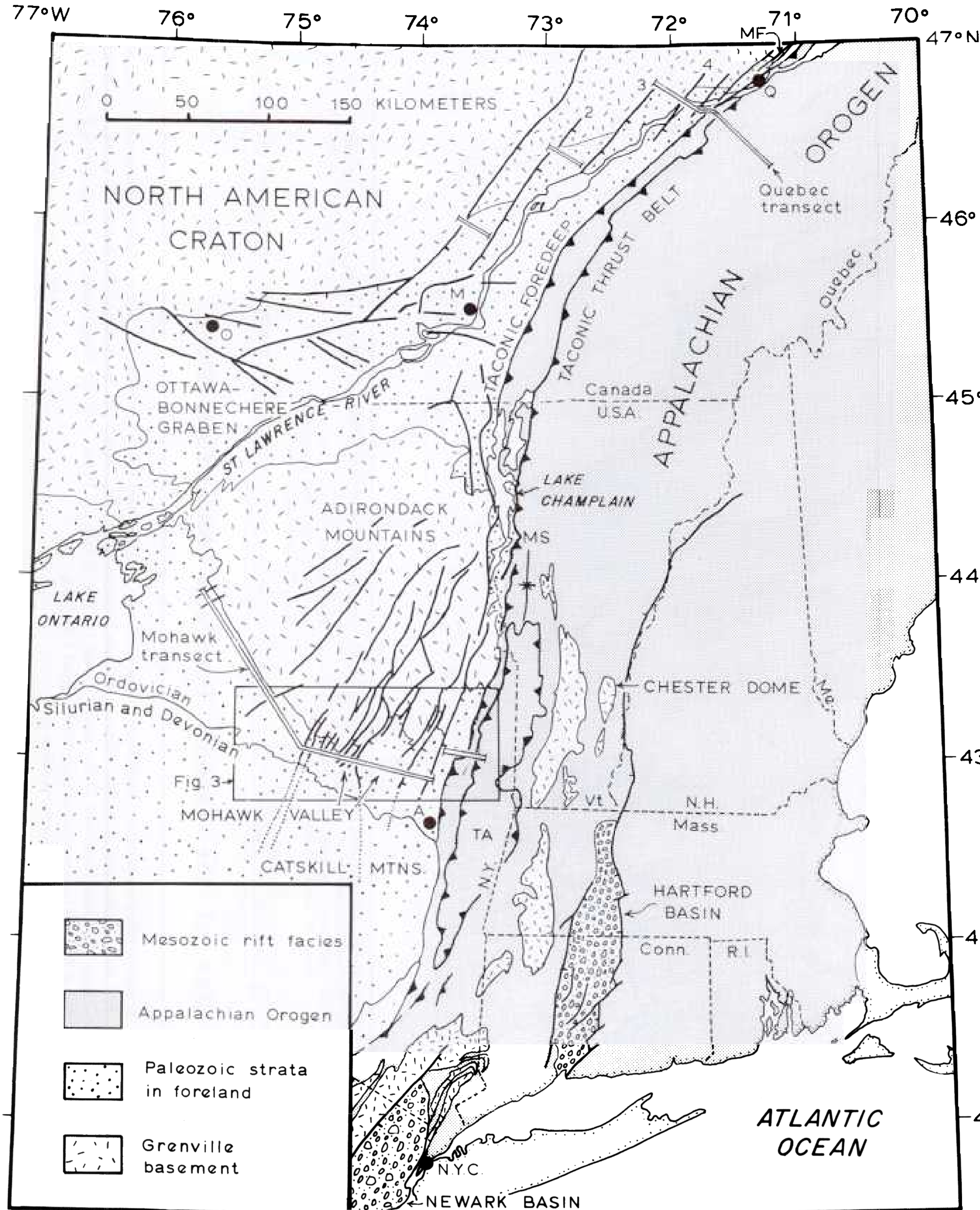

Figure 2. Map showing the Taconic fold-thrust belt and adjacent foreland in New York, Vermont, and Quebec, and locations of the Mohawk Valley and Quebec transects. Ordovician flexure-induced normal faults (heavy lines, barbed on downthrown side, dashed in subsurface of Catskill region) strike subparallel with contractional structures in the orogen. Fault nos. 1–4 in Canada are keyed to Table 2 and Figure 11B. Abbreviations: A, Albany; M, Montreal; MF, Montmorency fault; MS, Middlebury Synclinorium; MV, Mohawk Valley; NYC, New York City; O, Ottawa; Q, Quebec City; TA, Taconic Allochthon. Rectangle shows area of Figure 3.

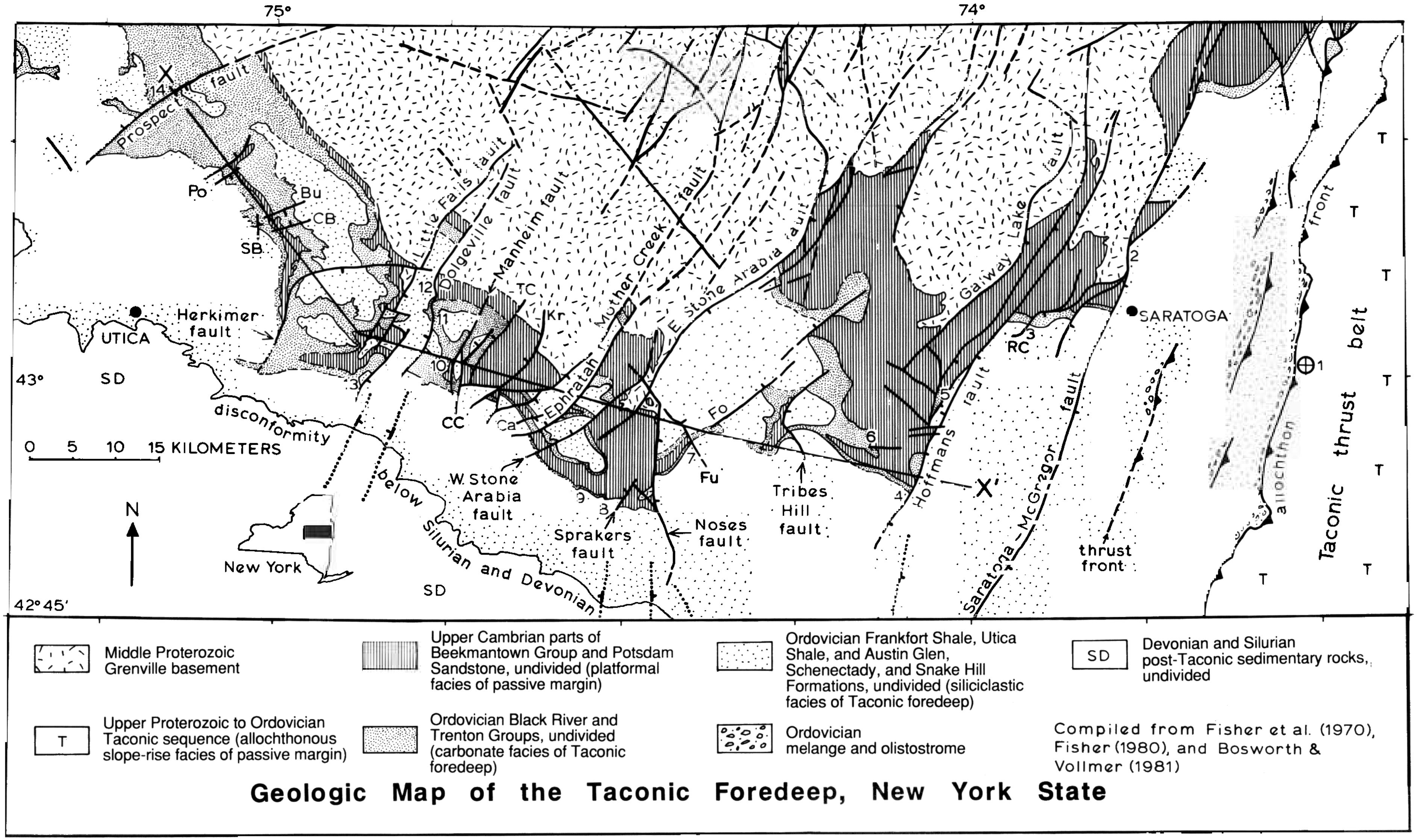

Figure 3. Generalized geologic map of the Taconic foredeep, eastern New York State. Ordovician normal faults are located west of, and strike subparallel with, Ordovician thrusts, which crop out in the eastern part of the map area. Simplified from Fisher and others (1970), with modifications from Fisher (1980) in the area between and including the Manheim and Hoffmans faults; and from Bosworth and Vollmer (1981) for the position of frontal thrusts. The Mohawk Valley approximately follows the line of section X–X′. Abbreviations for minor faults: Po, Poland; SB, Shedd Brook; Bu, Buttermilk Creek; Ca, Caroga Creek; CB, City Brook; CC, Crum Creek; Kr, Kringsbrush; Fu, Fultonville; Fo, Fonda; RC, Rock City Falls.

lithospheric flexure. We coin the term "flexural extension" for extension resulting from bending of a lithospheric plate. Flexural extension of continental lithosphere has been neglected but is interesting for several reasons. (1) Geologic data from foredeeps complement those from analogous modern deep-sea trenches, where detailed structural observations are difficult and costly. (2) Flexure-induced normal faulting is clear evidence for bending of the lithosphere beyond its elastic limit, contrary to some mathematical models of lithospheric flexure at convergent plate boundaries. (3) The structural geology of normal faults bears on mechanical models of lithospheric flexure (for example, Price and Audley-Charles, 1983; Freund and others, 1980). (4) Flexural extension is quite a different tectonic process from rifting, because it involves simultaneous extension (at the top) and contraction (at the base) of a single vertical column of lithosphere. (5) In some foredeeps, flexure-induced normal faults and related structures helped control the distribution of hydrocarbons and (or) Mississippi Valley–type lead-zinc deposits.

The effects of flexural extension are well preserved in the Taconic collisional foredeep in the Mohawk Valley region of upstate New York (Figs. 2 and 3). Taconic-age (Ordovician) normal faults are the dominant map-scale structures in a >100-km-wide belt flanking the thrust belt. This paper presents a structural analysis of the Mohawk Valley, and a review of published data from other normal-faulted foredeeps: the Taconic foredeep in the Champlain Valley and Quebec, the Arkoma Basin (the foredeep of the Ouachita orogen), and Timor Trough–Aru Basin (the active foredeep of the Australia–Banda Arc collision). Normal faults in these foredeeps are unlike those in Figure 1b or 1c. We present evidence that flexural extension is characterized not by formation of a series of independent tension cracks or faults that die out at depth, but instead by development of a systematic network of gently rotational normal faults in the upper part of the crust (Bradley, 1987). We suggest that these faults are linked at depth along a low-angle detachment or shear zone where fault displacement and block rotation are accommodated.

TABLE 1. NORMAL FAULTS IN THE TACONIC FOREDEEP, MOHAWK VALLEY, NEW YORK

Fault	Distance to thrust front (km)*	Throw (m)†	Polarity	Cumulative heave (Σh) (m)§
W. Carthage	208	20	Synthetic	12
Stony Creek	202	20	Synthetic	23
Lowville	182	~20	Synthetic	35
Prospect**	127	58	Reverse	
Poland NW	116	34	Antithetic	54
Poland SE	115	18	Synthetic	65
Shedd Brook	110	6	Synthetic	68
Buttermilk Creek	108	9	Synthetic	73
City Brook	106	11	Antithetic	80
Herkimer	100	21	Synthetic	92
Little Falls A	89	12	Antithetic	99
Little Falls B	88	24	Antithetic	113
Little Falls C (main)	86	268	Synthetic	268
Dolgeville	83	37	Antithetic	289
Manheim	79	40	Synthetic	312
Crum Creek	78	61	Antithetic	347
Timmerman Creek	76	40	Antitethic	370
Kringsbush	72	37	Antithetic	391
Mother Creek	68	137	Synthetic	470
Ephratah	64	⩾162	Antitheric	564
Stone Arabia W	61	146	Antithetic	648
Stone Arabia E	58	24	Synthetic	662
Noses	54	296	Synthetic	833
Fultonville	51	⩾31	Synthetic	850
Tribes Hill	38	69	Synthetic	890
Hoffmans	23	381	Synthetic	1,110
Saratoga	8	137	Synthetic	1,189

*Measured to thrust front in parautochthonous flysch; the allochthonous thrust front (Logan's Line) is about 13.5 km farther east.

†Sources of map and stratigraphic data for estimates of displacement: Fisher (1980); Miller (1909); Dunn (1954); Cushing and Ruedemann (1914); Johnsen (1958, 1970); Kay (1953). ⩾ indicates that the throw is a minimum value, probably not much less than the actual amount. ~ indicates an estimate from an unsatisfactory data source.

§Heave calculated assuming all faults have 60° dips.

**Prospect fault has 58-m reverse offset but has a fault bend fold in the hanging wall consistent with original normal down-to-east (synthetic) displacement. We think that this fault first had normal offset, but that it was removed and net reverse offset imposed during some later episode.

FLEXURAL EXTENSION IN THE TACONIC FOREDEEP, MOHAWK VALLEY, NEW YORK

Regional Geology

The Mohawk Valley region is underlain by middle Proterozoic metamorphic rocks of the Grenville Province (Fig. 3). This North American crystalline basement is overlain by Upper Cambrian siliciclastics (Potsdam Sandstone), followed by Upper Cambrian and Lower Ordovician marine carbonates (Beekmantown Group), which most workers regard as the deposits of a thermally subsiding passive margin that originated during Late Proterozoic rifting (for example, Bird and Dewey, 1970). Two observations suggest that prior to the Taconic orogeny, the shelf edge lay far to the east of the present thrust front: (1) only a few hundred meters of lower Paleozoic platformal strata are present in the Mohawk Valley (Rowley, 1982a), and (2) tectonized Grenville basement and lower Paleozoic strata of possible platformal affinity occur as far east as the Chester Dome in eastern Vermont (for example, Stanley and Ratcliffe, 1985) (Fig. 2).

The Appalachian passive margin collided with a subduction complex at the leading edge of a magmatic arc (or arcs) during Ordovician time, resulting in the Taconic orogeny (Stevens, 1970; Rowley and Kidd, 1981; Stanley and Ratcliffe, 1985; Bradley and Kusky, 1986; Bradley, 1989; Figs. 4 and 5). In the area of Figure 3, the first sign of impending orogenesis was a brief episode of uplift and erosion of the carbonate bank. A number of workers have attributed the resulting interval of unconformities (Fig. 4) to the passage of a forebulge (that is, a flexural bulge, formed as a result of vertical loading of a quasi-elastic plate, rather than by buckling under horizontal compression) across the carbonate platform (Chapple, 1973; Rowley and Kidd, 1981; Jacobi, 1981; Bradley and Kusky, 1986; Lash, 1988). Sedimentological studies in comparable carbonate strata in Virginia and Newfoundland have supported the forebulge model (Mussman and Read, 1986; Knight and others, 1991). The unconformity is overlain by an upward-deepening carbonate succession, the Caradocian Black River and Trenton Groups (Fig. 4). Trenton carbonates grade upward and eastward into carbonate turbidites (Dolgeville facies) and then carbonaceous hemipelagites (Utica Shale). The Utica, in turn, grades upward and eastward into submarine-fan deposits (for example, Schenectady Formation and Frankfort Shale) composed of turbiditic graywackes and shales derived from orogenic sources to the east (Rowley and Kidd, 1981). The shales and graywackes were deposited in a cratonward-migrating foredeep. At the present eastern margin of the foredeep, immediately adjacent to and beneath the Taconic allochthon (a far-traveled thrust of late Proterozoic to Ordovician continental rise facies originally deposited next to the North American passive margin), the proximal fill of the foredeep consists of graywackes, shales, olistostromes, and mélanges derived therefrom (Bosworth and Volmer, 1981). The foredeep probably resembled the present-day Timor Trough (for example, Karig and others, 1987), which is about 1.5 to 4 km deep at the thrust front (Bowin and others, 1980). At the end of Caradocian time (Fig. 5) when the Taconic thrust front reached its final position (Bradley and Kusky, 1986), the carbonate platform edge (in water depths of 0 to 0.2 km) lay about 100 km to the west, near Utica. Hence, we infer that the outer slope of the foredeep had a regional eastward dip of 1° to 2°. The regional dip direction is confirmed by paleocurrents (Cisne and others, 1982).

After Taconic plate convergence ended, the foredeep simultaneously rebounded and filled with east-derived clastics comprising the Late Ordovician to Silurian Queenston-Bloomsburg "delta." By the time platform carbonates of the Helderberg Group were deposited near sea level in Early Devonian time, the Trenton-Utica contact had rebounded to subhorizontal across all but the easternmost part of the foredeep. An Acadian foreland basin (filled by sediments of the Catskill "delta") reoccupied the site of the Taconic foredeep during a Middle and Late Devonian episode of flexural subsidence resulting from collision of the Avalon terrane, outboard of the Taconic arc(s) (Bradley, 1983).

Except in the extreme east and near normal faults, Paleozoic strata shown in Figure 3a are subhorizontal. Stratigraphic contacts in the Mohawk Valley dip regionally south-southwest about 1° (about 20 m/km or 100 ft/mi). The regional dip is probably due to both late Paleozoic loading of the Alleghanian thrust belt in Pennsylvania to the south, and to Neogene domal uplift of the Adirondack Mountains to the north (Isachsen and others, 1981) (Fig. 2). The regional dip produces essentially no apparent dip along the line of section near the Mohawk Valley (Fig. 3); the occurrence of westerly bedding dips of 1° to 2° in parts of the profile is due to normal faulting.

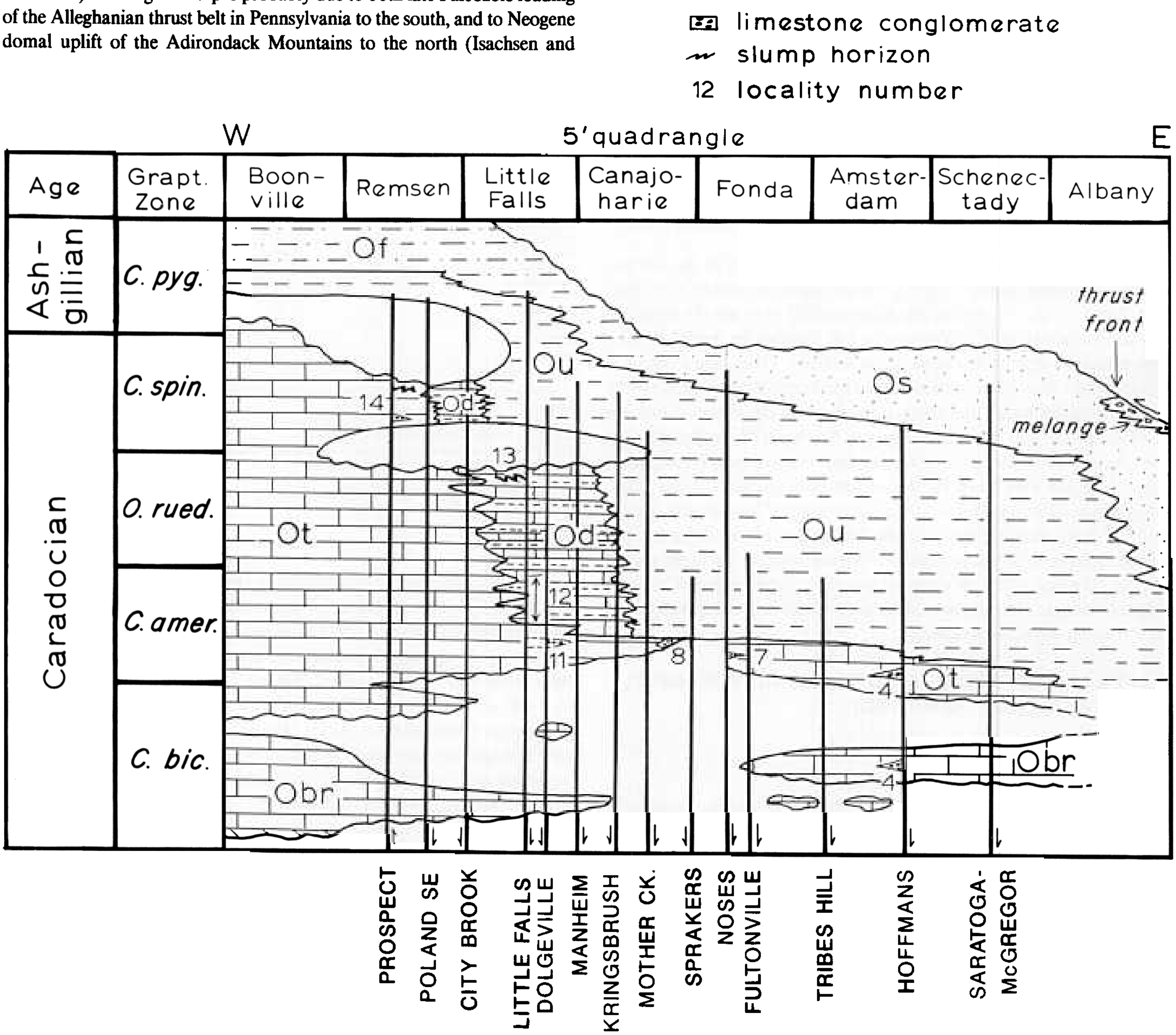

Figure 4. Middle and Upper Ordovician stratigraphy of the Mohawk Valley region, New York, adapted from Fisher (1977, 1979). Trenton (Ot) and Black River (Obr) Groups and Dolgeville Facies (Od) comprise the carbonate facies of the cratonic flank of the Taconic foredeep. Utica Shale (Ou), Frankfort Shale (Of), and Schenectady Formation (Os) comprise the upward- and eastward-coarsening siliciclastic fill of the foredeep. Caradocian facies belts migrated westward in response to plate convergence. Bold vertical lines represent Mohawk Valley faults and terminate upward at the time when motion ceased. A few closely spaced faults have been omitted for clarity. Fault deactivation was estimated from the stratigraphic position of the oldest undisplaced horizon, based on its elevation with respect to the base and top of the formation. Faults in the east were overlapped by foredeep siliciclastic rocks before those in the west; even the long-lived Saratoga-McGregor, Hoffmans, and Noses faults evidently stopped moving before minor faults in the west such as the City Brooks did. Numbers refer to localities mentioned in the text where evidence exists for synsedimentary faulting. The Hoffmans fault, in the east, is thought to have moved first; farther west, the Noses, Sprakers, and Little Falls faults began moving slightly later; still farther west, motion on the Prospect fault did not begin until considerably later, at about the time the Taconic thrust front reached its final position. Abbreviations for graptolite zones (from Riva, 1974, as modified by Finney, 1986): *N. grac.*, *Nemagraptus gracilis*; *C. bic.*, *Climacograptus bicornis*; *C. americanus*, *Corynoides americanus*; *O. rued.*, *Orthograptus ruedemanni*; *C. spin.*, *Climacograptus spiniferus*, *C. pyg.*, *Climacograptus pygmaeus*.

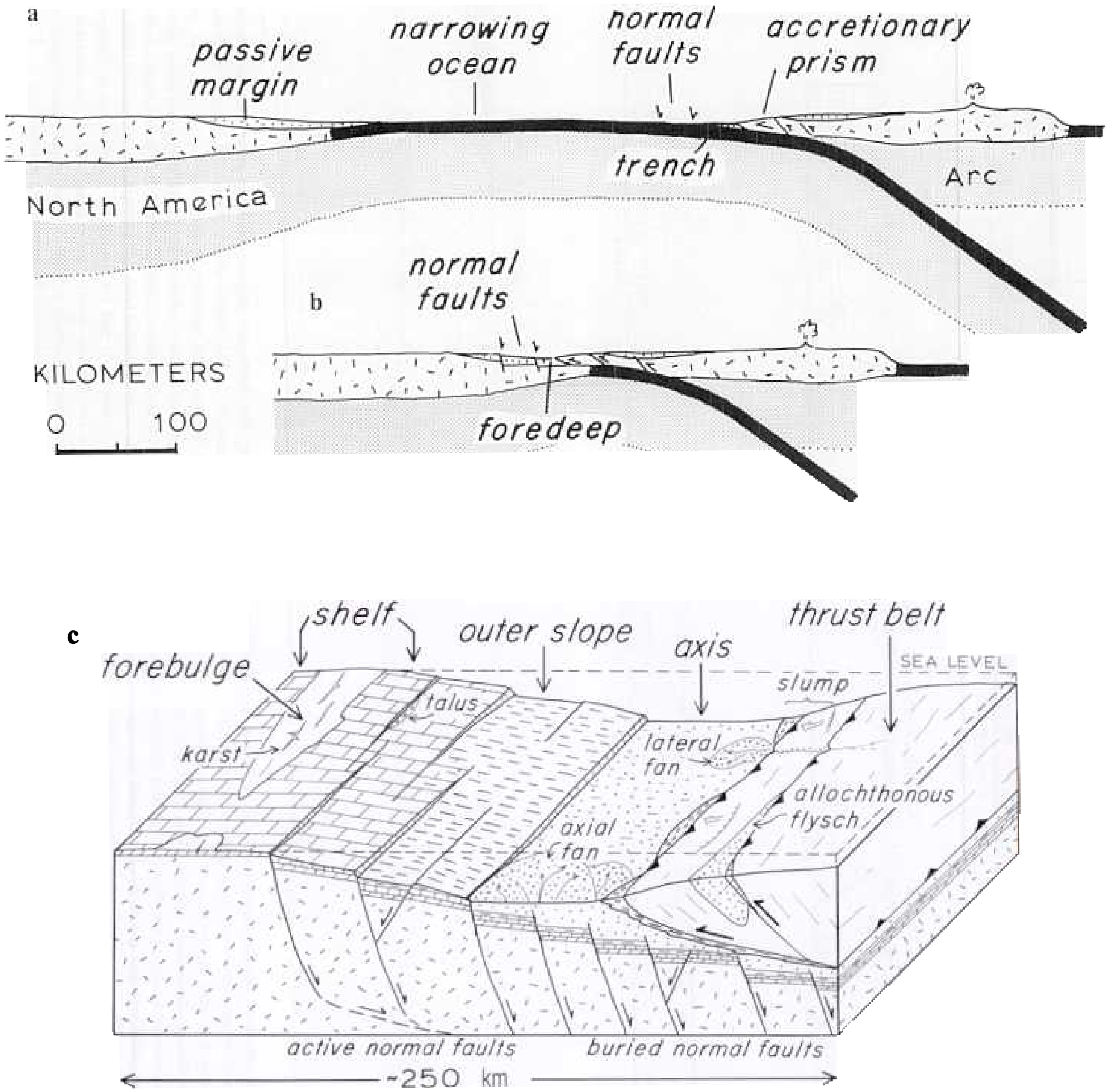

Figure 5. (a and b) Plate-tectonic model for the Taconic orogeny involving arc–passive margin collision. In 5a, the oceanic crust of the outer trench slope was presumably cut by normal faults; the passive margin passed through this regime of normal faulting as it approached the trench. Random dashes, continental crust; black, oceanic crust; gray, mantle lithosphere. (c) Schematic block diagram of the Taconic collisional foredeep, showing the distribution of facies belts and structural regimes shortly before plate convergence ended.

Normal Faults in the Mohawk Valley

Subsidence of the Taconic foredeep was accompanied by motion on normal faults[1], which mostly strike parallel with the thrust front (Figs. 2 and 3), except in the extreme west. The larger faults are spaced about 10–20 km apart and can be traced tens of kilometers along strike. From east to west, the most important are the Saratoga-McGregor, Hoffmans, Noses, and Little Falls faults. A majority dip and downdrop to the east (toward the orogen, here termed "synthetic faults"); west-dipping antithetic faults are also present, but are less numerous and have mostly smaller displacements (Fig. 6). Stratigraphic throws range from a few meters in the extreme west (>200 km from the thrust front), to a few

[1]An appendix and two figures describing in detail the structural geology of the normal-faulted Mohawk Valley region are available free of charge by requesting Supplementary Data 9127 from the GSA Documents Secretary.

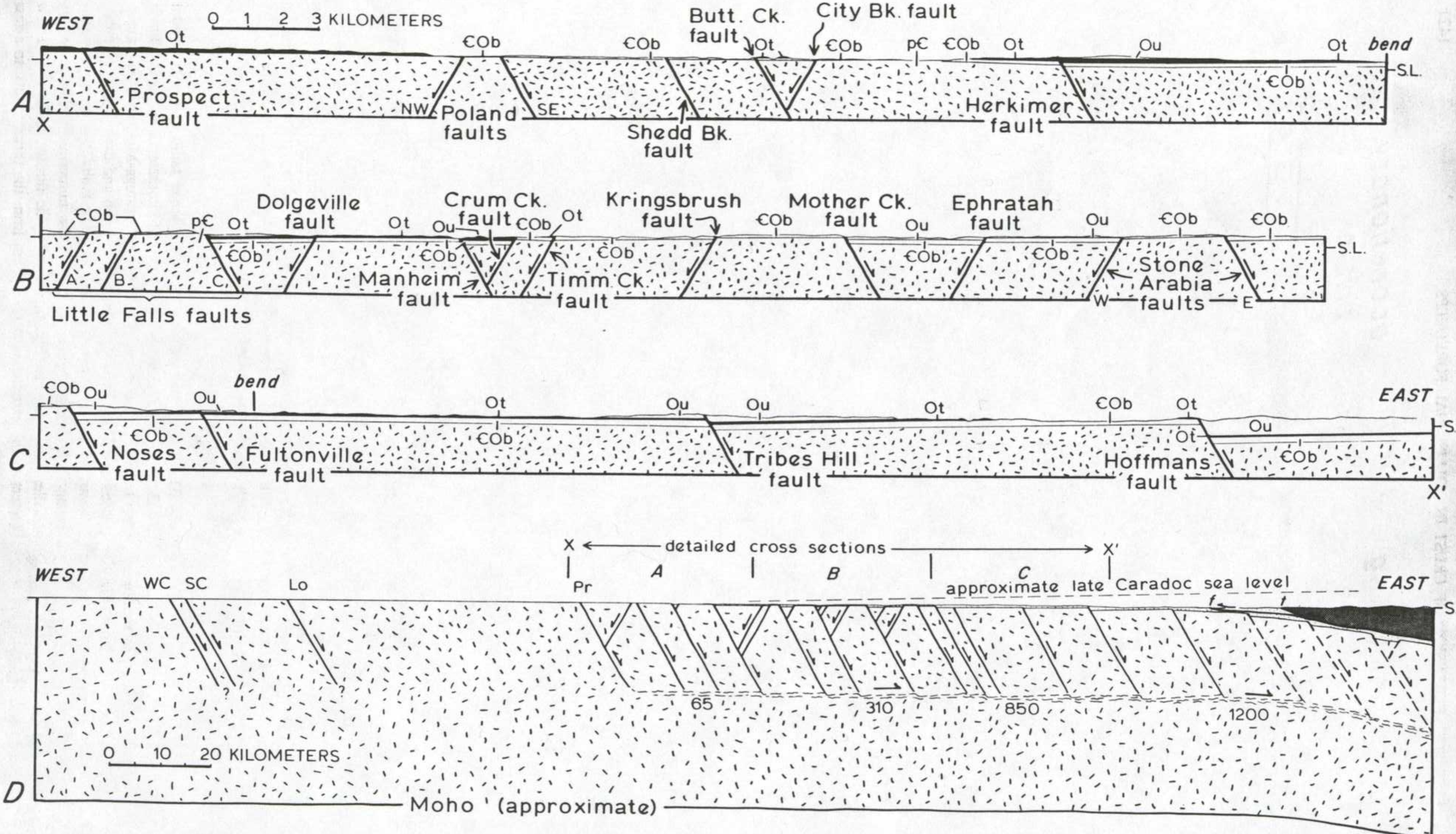

Figure 6. (A–C) West-east schematic structure section along line X–X′ in Figure 3. Faults that obliquely intersect the line of section are shown with true rather than apparent dips. The Shedd Brook fault has been projected onto the section a short distance along strike. Random dashes denote Grenville continental basement. Abbreviations for Paleozoic strata: COb, Potsdam Sandstone and Beekmantown Group; Ot, Black River and Trenton Groups; Ou, Utica Shale. Compiled from published and unpublished 1:24,000 geologic maps listed in the legend of Fisher and others (1970), with modifications from Fisher (1980) and our own observations. (D) Composite cross section which summarizes detailed sections A, B, and C, and also extends to the east and west along the line shown in Figure 2. WC, West Carthage fault; SC, Stony Creek fault; Lo, Lowville fault. Eastward dips of most of the normal faults and gentle westward dips of Paleozoic strata together imply a rotational normal fault geometry. The faults are shown as planar structures intersecting a low-angle detachment or zone of ductile simple shear, which we suggest is probably 15–20 km deep; alternatively, a listric fault geometry is also possible (see Fig. 15D, inset). Any space problems where high-angle faults intersect the postulated basal shear zone are assumed to have been accommodated by some combination of ductile and brittle deformation. Numbers at the base of the normal faulted layer are estimated accumulated horizontal displacement, in meters (from Table 1). F and TA denote the thrust front in Ordovician flysch and the leading edge of the Taconic allochthon (Logan's Line), respectively. No attempt has been made to show structural details in the thrust belt (shaded area), such as the degree of involvement of basement and parautochthonous cover in thrusting.

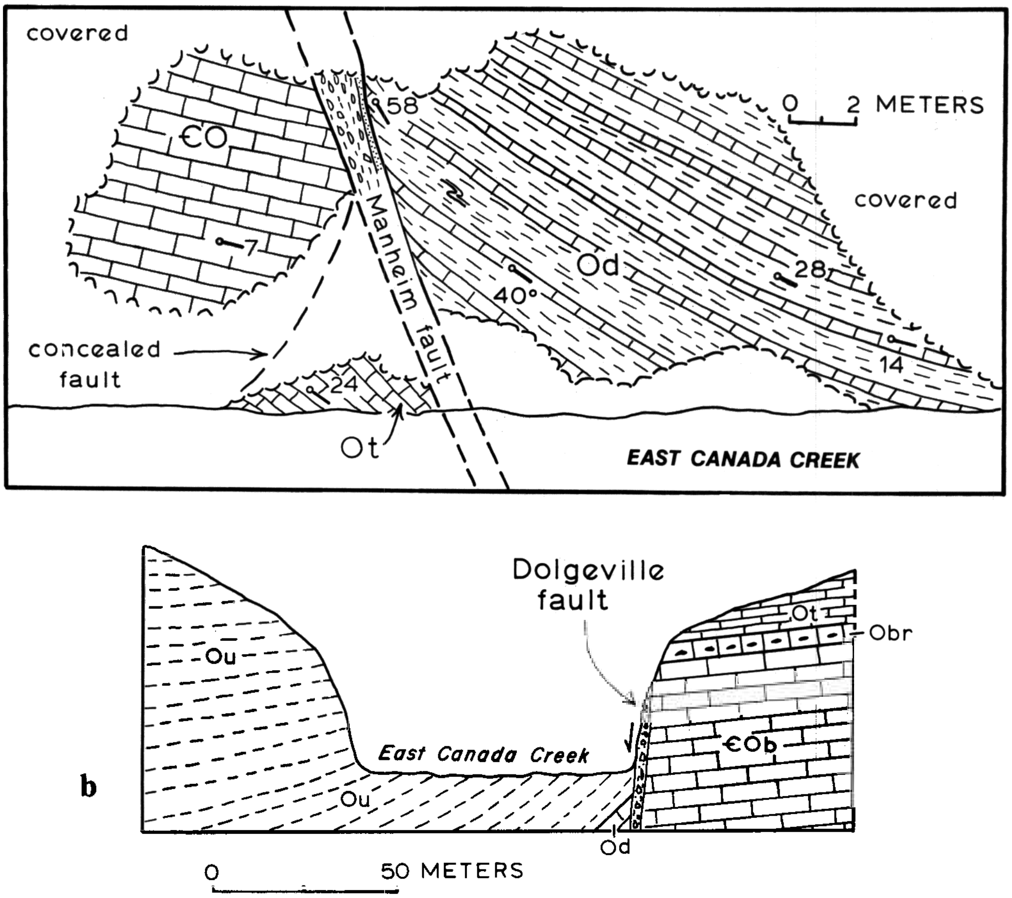

Figure 7. Cross sections through normal faults in the Taconian foredeep. Rock-unit symbols as in Figures 4 and 6. (a) Cross-sectional sketch from a photograph of the Manheim fault along East Canada Creek, viewed toward the north-northeast. Dip symbols show true dips as measured at outcrop. Dolostones of the Beekmantown Group (footwall, left) are downdropped against shales and deep-water carbonates of the Dolgeville facies (hanging wall, right). The fault zone, which dips 77°E, is a well-cemented carbonate breccia 1–2 m thick, intruded by a 25-cm-thick kimberlite dike (stipple); a nearby dike of the same composition yielded an Early Cretaceous Rb/Sr age of 130 Ma (Fisher, 1980). Bedding in both fault walls steepens toward, and dips in the same direction as, the fault; this effect is more pronounced on the downthrown side. A minor, stratabound slump fold ("Z") in downthrown strata near the fault may record syn-Dolgeville tilting of the sea floor as a consequence of contemporaneous faulting. A fault-bounded sliver of lower Trenton Group limestones crops out at water's edge. Unshaded area is vegetated. (b) Cross section of Dolgeville fault, after Cushing (1905). View toward the north-northeast.

hundred meters on normal faults nearer the thrust front (Table 1). The normal faults of the Mohawk Valley have been recognized since the first Geological Survey of New York (Vanuxem, 1842); regional mapping and local studies have established the map pattern, sense and amount of displacement, structural style, and some age constraints (Darton, 1895; Cushing, 1905; Megathlin, 1938; Fisher, 1980; Cisne and others, 1982; Bosworth and Putman, 1986). We compiled the cross section shown in Figure 6 from 1:24,000 scale published and open-file geologic quadrangle maps, supplemented by our outcrop observations. Stratigraphic thicknesses are from the geologic maps and from Fisher (1977). Like previous workers, we determined stratigraphic throw from map elevations of one or more subhorizontal contacts on either side of each fault. Contacts suitable for this purpose include the Grenville-Potsdam, Beekmantown–Black River, and Trenton-Utica.

Fault dips are not well constrained but average about 60°. Only one map-scale normal fault in the study area is well exposed, the Manheim, which dips 77°E (Fig. 7a). The Prospect fault dips 58°; it now has reverse offset but probably originated as a normal fault (see Table 1 and appendix). Several other fault exposures mentioned by Cushing (1905), Megathlin (1938), and Fisher (1980), either no longer exist or are degraded. At locality no. 12, the antithetic Dolgeville fault dips about 85°, judging from an illustration by Cushing (1905) that has been redrawn as Figure 7b. Minor normal faults with displacements of a few centimeters to a few meters are present at a number of the larger exposures in the Mohawk Valley (Fig. 8a), and in some cases, dips are quite variable along individual faults, suggesting that individual dip readings at single outcrops may not be representative. Measured dips range from 24° to 87° on synthetic minor faults, and 30° to 83° on antithetic minor faults, but most dips exceed 55° (see appendix). Comparable major and minor normal faults exposed in the Champlain Valley and Quebec (see below) typically have 60° to 65° dips. Accordingly, we assumed 60° dips on all faults when constructing the cross section.

Map-scale faults are characterized by a breccia of angular fragments from one or both fault walls. Such fault breccias occur along the McGregor (Bosworth and Putman, 1986), Manheim, and Dolgeville fault traces (localities 2, 10, and 12 in Fig. 3). Sulfide occurrences in the Manheim and Dolgeville fault zones (Fig. 7; Vanuxem, 1838; Cushing, 1905) reveal that these faults acted as fluid conduits.

Bedding near most faults shows pronounced normal "drag" on the downdropped block, steepening from subhorizontal a few tens of meters from the fault, to 60° or more, thereby forming an asymmetric syncline (Fig. 7). Downdropped strata next to the Dolgeville fault dip as steeply as 64°; beds next to the Manheim fault dip 58°, and beds next to a minor antithetic fault near the Ephratah fault dip 46° (our observations). Next to the Saratoga-McGregor fault, Isachsen and others (1981, p. 298) reported easterly dips up to 60°. Megathlin (1938, p. 105) reported average of bedding dips 60° and dips as great as 70°, but he did not mention localities; he suggested that fault dips cannot be much steeper than the maximum dip of "dragged" beds. In all cases, whether the fault is synthetic or antithetic, downdropped beds dip away from the fault. Normal "drag" is barely

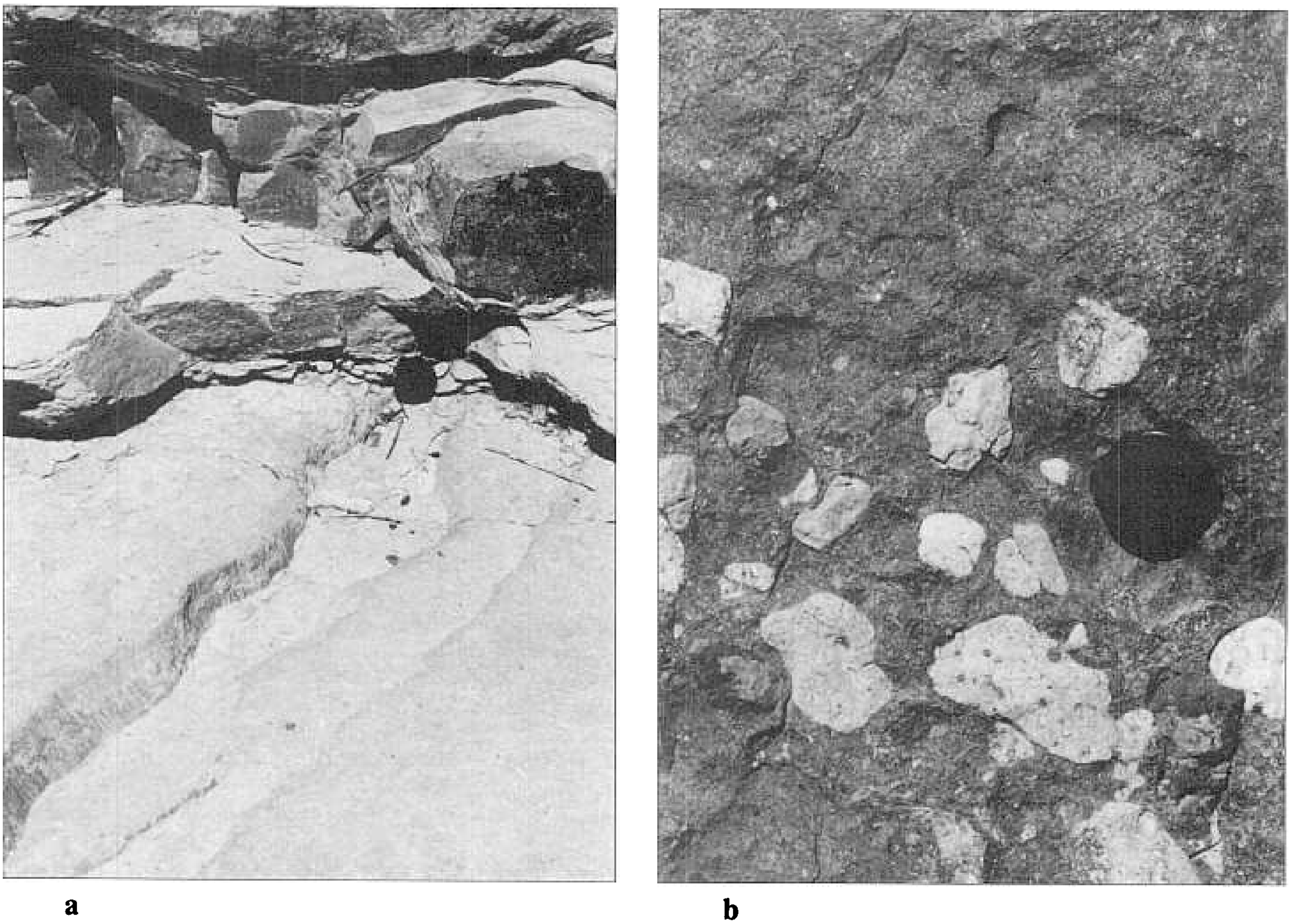

Figure 8. (a) View toward the north-northeast of minor synthetic (right-dipping) and antithetic normal faults cutting Black River Group carbonates at Inghams Mills. Slickenlines plunge down the dip of fault planes. Lens cap for scale. (b) Limestone-matrix conglomerates in the Trenton Group at Inghams Mills (Fig. 3), between the Dolgeville and Manheim faults. The conglomerates, which contain clasts of subjacent Cambrian and Ordovician carbonates and Proterozoic basement gneisses, are inferred to have been shed from a nearby fault scarp, probably the Little Falls fault.

evident on the upthrown side of normal faults. The Manheim fault is typical (Fig. 7a); at locality 10, footwall beds steepen within a few meters of the fault to a maximum dip of just 7°.

The block-faulted structure of the Mohawk Valley region has been recognized in the parautochthonous passive margin sequence of the Middlebury Synclinorium (Fig. 2), structurally below the Taconic Allochthon (Zen, 1967, 1970; Rowley, 1982b). Stratigraphic evidence reviewed by Zen (1967, p. 41–42; 1970, p. 133–134) suggests that normal faulting occurred during deposition of the Caradocian Ira Formation, which is lithologically and paleogeographically akin to the Utica Shale. Because they exist to the east and west, there is every reason to suspect that Caradocian-age normal faults also exist beneath the Taconic allochthon, as is shown schematically in Figure 6d. High-resolution industry seismic profiles in this key area were run in the early 1980s, but they remain confidential. A test well near the allochthon front at Willard Mountain (Fig. 3, locality no. 1) (Columbia Gas Co., 1983) intersected the Trenton-Utica contact at a depth of 1,470 m below sea level, indicating that the autochthon descends to considerable depth beneath the thrust belt. It seems likely that downdropping on subsurface normal faults is at least in part responsible.

Figure 6D illustrates the relationships between normal and thrust faulting in the study area, and a speculative interpretation of fault geometry at depth. The eastern part of the cross section shows a striking predominance of down-to-east faults, and associated gentle (typically $<1°$) westerly bedding dips, as recognized by Megathlin (1938). These features, and similar observations from other normal faulted foredeeps, require that the normal faults are gently rotational with respect to the enveloping surface of the flexed plate, and they suggest to us that at depth, the high-angle faults intersect a low-angle detachment or zone of distributed ductile simple shear. The rationale for this cross-sectional fault geometry is developed more fully in the section "Generalizations on Flexural Extension." The estimated position of a Carodocian sea-level datum, which now dips 1° to 2° west, is also shown; this was drawn by assuming that the foredeep bathymetry was comparable to that of the present-day Timor Trough (Veevers and others, 1978). Post-Taconic rebound has elevated the former foredeep axis above present sea level, but at the time of extension, the enveloping surface of the North American plate would have dipped 1° to 2° east (present coordinates).

Age of Normal Faulting

Chadwick (1917) postulated a "relation of the Adirondack-Mohawk step-faults to the great charriage movements of New England over eastern New York, which, by overloading, may have depressed successive fragments of the overridden area." Chadwick's hypothesis was never elaborated in a full-length paper and was dismissed by most of his contemporaries and many later workers, but a variety of evidence reviewed below strongly supports his hypothesis that the normal faulting was primarily syn-Taconic in age (Cisne and others, 1982; Bradley and Kusky, 1986; Hay and Cisne, 1989; Mehrtens, 1989a).

In the southern part of the area shown in Figure 3, all of the normal faults die out within Ordovician strata, an observation for which two alternative explanations are possible: (1) the faults actually end before reaching the erosional edge of Silurian and Devonian rocks; or (2) the faults exist at depth, buried by Ordovician foredeep deposits. Megathlin (1938, p. 119) favored the first hypothesis; he related the faults to post-Taconic "relaxation." On the basis of well logs, Rickard (1973) has since recognized horst-and-graben structure in sub-Silurian strata south of the Mohawk Valley, indicating that faults continue southward at depth. Isachsen and McKendree (1977) traced the Little Falls, Noses, and Hoffmans faults in the subsurface tens of kilometers to the south of the post-Ordovician unconformity (Figs. 2 and 3). Although some of the southward decrease in throw might be due to hinged differential uplift of portions of the Adirondacks, the above observations require at least a dominant component of Caradocian growth faulting.

Several faults are flanked by tongues of Caradocian-age conglomerate within platform carbonates of the Black River and Trenton Groups. Just west of the Hoffmans fault (locality no. 4; Figs. 3 and 4), limestone-matrix conglomerates within the Amsterdam Limestone (Black River Group) contain clasts of the underlying Beekmantown Group (Fisher, 1980, p. 18). Similar conglomerates also occur in the overlying Larrabee Member of the Glens Falls Limestone (lower Trenton Group) at localities no. 4 and no. 5; clasts in these younger conglomerates were derived from both the Black River and Beekmantown Groups. The localized occurrence of conglomerates near the Hoffmans fault, and their lateral gradation into nonconglomeratic limestones at Manny Corners just 6 km across strike to the west (locality no. 6; Park and Fisher, 1969), suggest that they were shed from an uplifted fault block that had risen above sea level from a previously smooth carbonate platform (Bradley and Kusky, 1986, p. 675). Conglomerates in the lower Trenton Group just east of the Noses fault (Van Wie Creek, locality no. 7) contain clasts of Potsdam-type sandstone and Beekmantown-type dolostone (Cisne and others, 1982, p. 235). Cisne and others (1982) suggested that the conglomerates were derived from the upthrown footwall of the Noses fault, where Utica Shale directly overlies Beekmantown Group, and the Trenton Group is absent. South of section X–X′, the Trenton Group reappears on the downdropped west side of the antithetic Sprakers fault. At locality no. 8 (Flat Creek), the Trenton Group includes boulder conglomerates derived from Beekmantown Group dolostones (Kay, 1937, p. 264); however, at locality no. 9 (Canajoharie Gorge), 5 km across strike to the west, conglomerates are absent. Hence, the Noses and Sprakers faults bounded a horst during Trenton deposition (Cisne and others, 1982, p. 242). On the upthrown block between the Manheim and Dolgeville faults at locality no. 11 (Inghams Mills), conglomerate in the Trenton Group contains clasts of Beekmantown Group and Grenville gneiss (Fig. 8a). If the clasts were derived from a nearby fault scarp, the most likely place is from the upthrown western side of the

TABLE 2. NORMAL FAULTS IN THE TACONIC FOREDEEP, ST. LAWRENCE VALLEY, QUEBEC

Fault*	Distance to thrust front (km)†	Throw (m)§	Polarity	Cumulative heave (Σh) (m)**
(1) Ste. Sophie-Ste. Julienne	84	351	Synthetic	203
(2) St. Cuthbert	59	360	Synthetic	410
(3) St. Prosper	36	236	Synthetic	547
(4) Deschambault	20	227	Synthetic	678
(5) Jacques-Cartier River	9	375	Synthetic	894
(6)	8	50	Synthetic	923
(7)	6	125	Synthetic	995
(8) Neuville	5	440	Synthetic	1,249
(9)	3	375	Synthetic	1,466
(10)	2	80	Synthetic	1,512
(11)	0	95	Synthetic	1,567
(12)	−3	200	Synthetic	1,682
(13)	−8	220	Synthetic	1,809
(14)	−11	190	Synthetic	1,919
(15)	−14	190	Synthetic	2,029
(16)	−21	940	Synthetic	2,572
(17)	−24	1,000	Synthetic	3,149
(18)	−31	780	Synthetic	3,599
(19)	−35	595	Synthetic	3,943
(20)	−38	750	Synthetic	4,376
(21)	−42	?	Synthetic	?

*Numbers refer to faults in Figure 2 (nos. 1–4), and those intersected by the SOQUIP profile in Figure 10a (nos. 5–21). On the SOQUIP profile (St. Julien and others, 1983), fault no. 6 rather than fault no. 5 was identified as the Jacques–Cartier River fault, and fault no. 11 rather than fault no. 8 was identified as the Neuville fault.

†Measured to thrust front in parautochthonous flysch; the allochthonous thrust front (Logan's Line) is about 10.9 km farther southeast. Negative values are behind the thrust front.

§Sources of map and stratigraphic data for estimates of displacement: Faults 1–4, Clark and Globensky (1973, 1975a, 1975b, 1976a, 1976b); faults 5–21, St. Julien and others (1983).

**Heave calculated assuming all faults have 60° dips.

Little Falls fault, about 5–6 km to the west, where basement rocks are locally exposed today.

Prominent slump horizons occur within the Trenton Group near two faults and record synsedimentary tilting of the sea floor. Just west of the main Little Falls fault, gently west-dipping, deep-water carbonates of the Caradocian Dolgeville facies contain west-directed slump folds (locality no. 13) (Fisher, 1979); this is evidence for cratonward tilting of the Little Falls footwall block during Dolgeville sedimentation. Two slump horizons in the Trenton Limestone at locality no. 14 probably record activity on the Prospect fault.

The synthetic Little Falls fault (throw, 268 m) and the antithetic Dolgeville fault (throw, 40 m; Fig. 7b) together define a graben containing strata as young as Utica shale downdropped between rocks as old as Grenville basement. In addition to the Dolgeville-age slump folding on the Little Falls footwall block, four other lines of evidence also suggest that these faults were active during Caradocian time. (1) The main Little Falls and Dolgeville faults do not cut the post-Ordovician unconformity to the south, but instead die out within foredeep clastic rocks. (2) The lower part of the Dolgeville facies (*Corynoides americanus* zone) is thicker within the graben than to the east or west, as revealed by the spacing between correlative ash layers (Cisne and others, 1982, p. 232 and 236). (3) The youngest anomalously thick strata bear a deeper-water fauna, dominated by the trilobite genus *Triarthrus*, compared with coeval strata outside the graben, which are dominated by the shallower-water trilobite genus *Flexicalymene* (Cisne and others, 1982, p. 237). (4) Carbonate turbidites of the Dolgeville facies thicken systematically toward the Dolgeville fault at locality 12. We measured two detailed sections of an identical stratigraphic interval, one about 100 m and the other about 50 m from the fault. The sections thicken from 11 to 12 m, respectively.

A compilation of all available evidence confirms Bradley and Kusky's (1986) finding, that the locus of normal faulting migrated cratonward during Caradocian time (Fig. 4). The larger faults in the east were active longer than in those in the west and accumulated greater

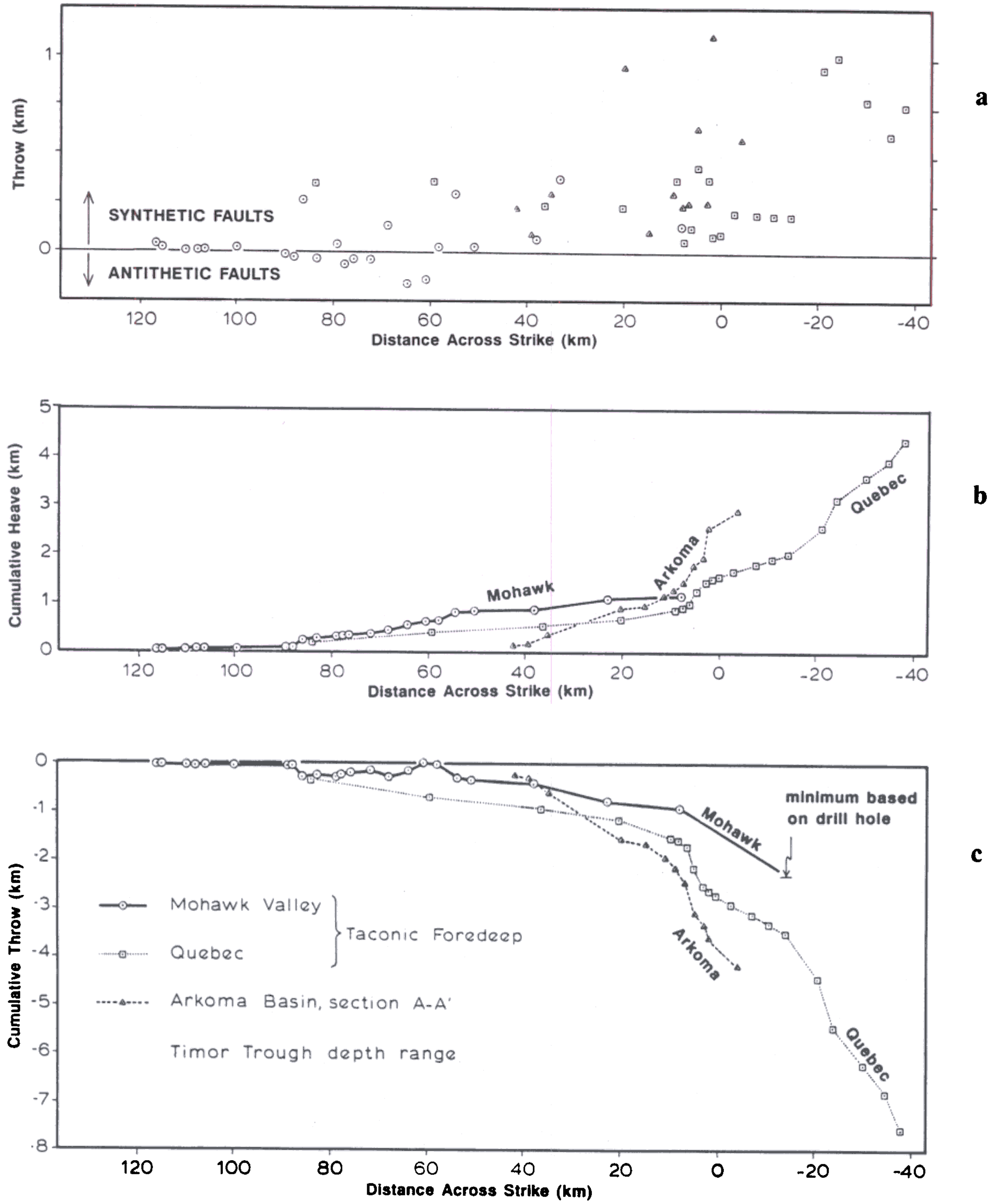

Figure 9. Three variables plotted against position (distance across strike) for normal faults from foredeep sections in the Mohawk Valley, Quebec, and Arkoma basin (Tables 1–3). Key to symbols is in Figure 9c. Along the horizontal axis, 0 km corresponds to the present thrust front. All Mohawk Valley curves ignore the Prospect fault, for reasons given in Table 1. (a) Throw versus position; note the predominance of synthetic faults and small throws in the distal part of each foredeep. Although antithetic faults are reasonably abundant in the Mohawk Valley, they have mostly smaller throws than do the synthetic faults. (b) Cumulative heave versus position. (c) Cumulative throw versus position. Displacements are plotted as negative for synthetic faults, and as positive for antithetic faults. Curves for all three sections show the same general shape that results from stepwise descent of the underthrust plate beneath the fold-thrust belt. In contrast, a flexed plate cut by symmetrical horsts and grabens (as in Fig. 1C) would plot as horizontal sawtooth line centered on the x-axis. Comparison with a range of bathymetric profiles across the Timor Trough–Aru basin (Bowin and others, 1980) reveals that for each of the three ancient foredeeps, normal faulting was capable of producing water depths like those in the best available modern analogue.

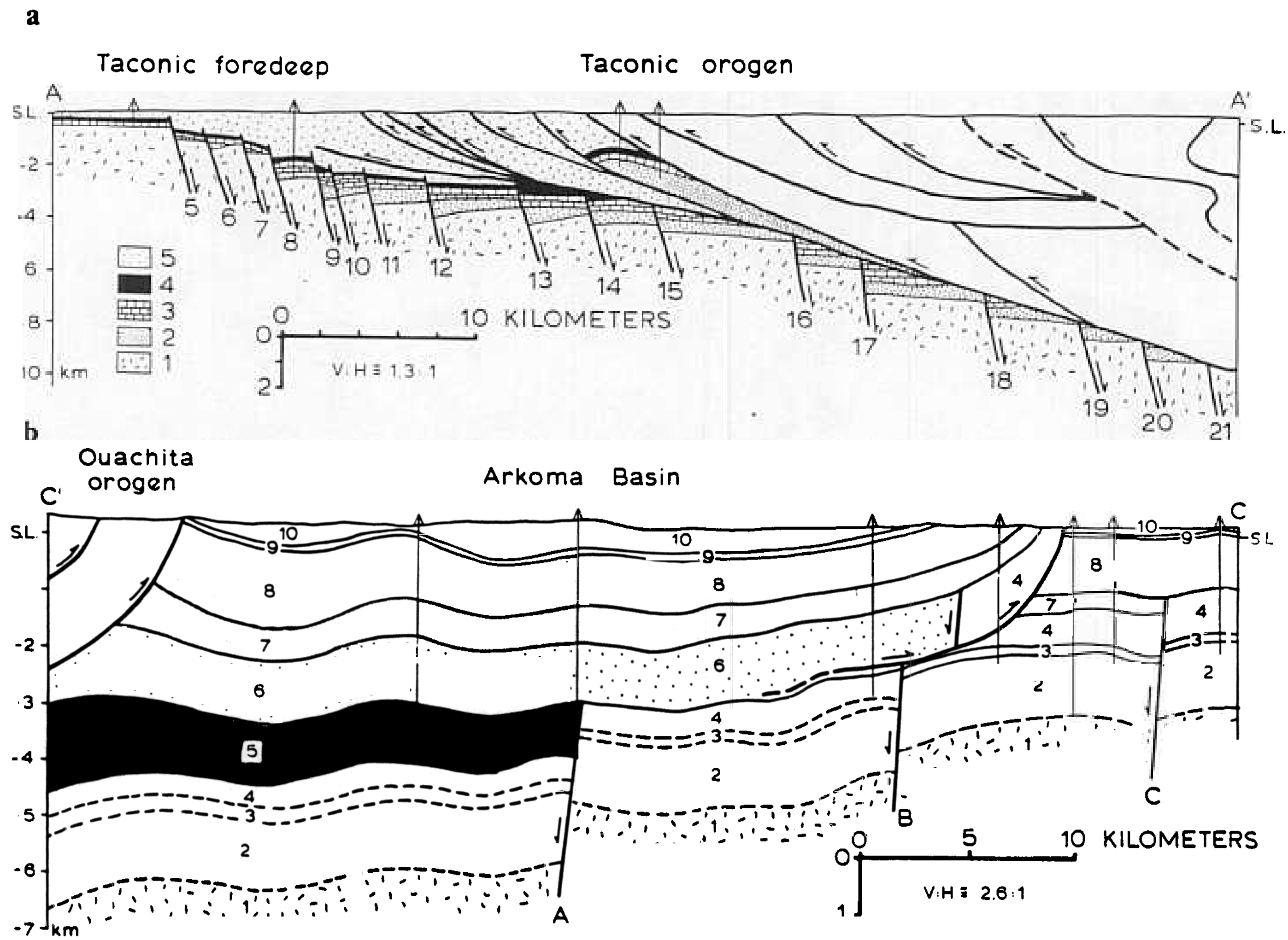

Figure 10. (a) Interpretation of a SOQUIP seismic reflection profile along the northern leg of the Quebec section in Figure 2, from St. Julien and others (1983). Unit 1, Grenville basement; unit 2, lower Paleozoic clastic rocks; unit 3, lower Paleozoic carbonate rocks, largely equivalent to Caradocian Trenton Group; unit 4, Caradocian Utica Shale; unit 5, Caradocian to Ashgillian flysch. Numbers on faults correspond with those in Table 2. (b) Cross section B–B′ through the Arkoma Basin, from Buchanan and Johnson, (1968). Unit 1, Precambrian basement; unit 2, Cambrian through Mississippian strata, undivided; unit 3, Morrow Formation; unit 4, lower part of Atoka Formation; units 5–7, middle part of Atoka Formation; unit 8, upper part of Atoka Formation; unit 9, Hartshorne Sandstone; unit 10, McAlester and Savannah Sandstones. Units 3–10 are Pennsylvanian. The distribution and thickness of mid-Atokan foredeep siliciclastic rocks (units 5–7) was controlled by normal faults, whereas older and younger strata are unaffected. A bedding-parallel thrust offsets normal fault B.

displacements (Fig. 9a). The Taconic allochthon reached its final position during late Caradocian time (age of wildflysch at the thrust front; see Fig. 4 and discussion and references in Bradley and Kusky, 1986, p. 677). Age relationships in Figure 4 indicate that in the Mohawk Valley region, the carbonate platform was uplifted at a forebulge and then cut by normal faults. While normal faulting accompanied foredeep subsidence in the west, the North American plate was being subducted farther east.

Evidence in the Mohawk Valley for normal faulting before or after the Taconic Orogeny is not compelling. It is reasonable to suspect that the area of Figure 3 may have been mildly extended during late Proterozoic rifting that led to opening of the Appalachian-Caledonian ocean, as Fisher (1977) showed schematically in his Plate 5. The only evidence for such extension in this region, however, is the occurrence in the Adirondacks of diabase dikes which cut Grenville basement but not the Cambrian Potsdam Sandstone, and which have yielded K-Ar ages in the range of 588–542 Ma (Isachsen and others, 1988). The total absence of late Proterozoic rift facies in the Mohawk Valley seems to preclude any *major* normal fault motion during that time. Similarly, there is no clear evidence, such as stratigraphic thickening or facies changes across faults, that normal faulting occurred during Cambrian to Early Ordovician passive margin subsidence. The most conspicuous extensional structures along the Eastern Seaboard are the product of Mesozoic rifting that led to opening of the Atlantic. The area of Figure 3, however, lies entirely inboard of the Mesozoic normal faults that bound the Newark and Hartford basins (Fig. 2). If Mesozoic faults were present in the Mohawk Valley, they would offset the post-Ordovician unconformity; instead, this unmistakable contact is essentially undisturbed (Figs. 2 and 3), except by Devonian or younger thrusts in the extreme east (Marshak, 1986). Possible evidence for Cenozoic mo-

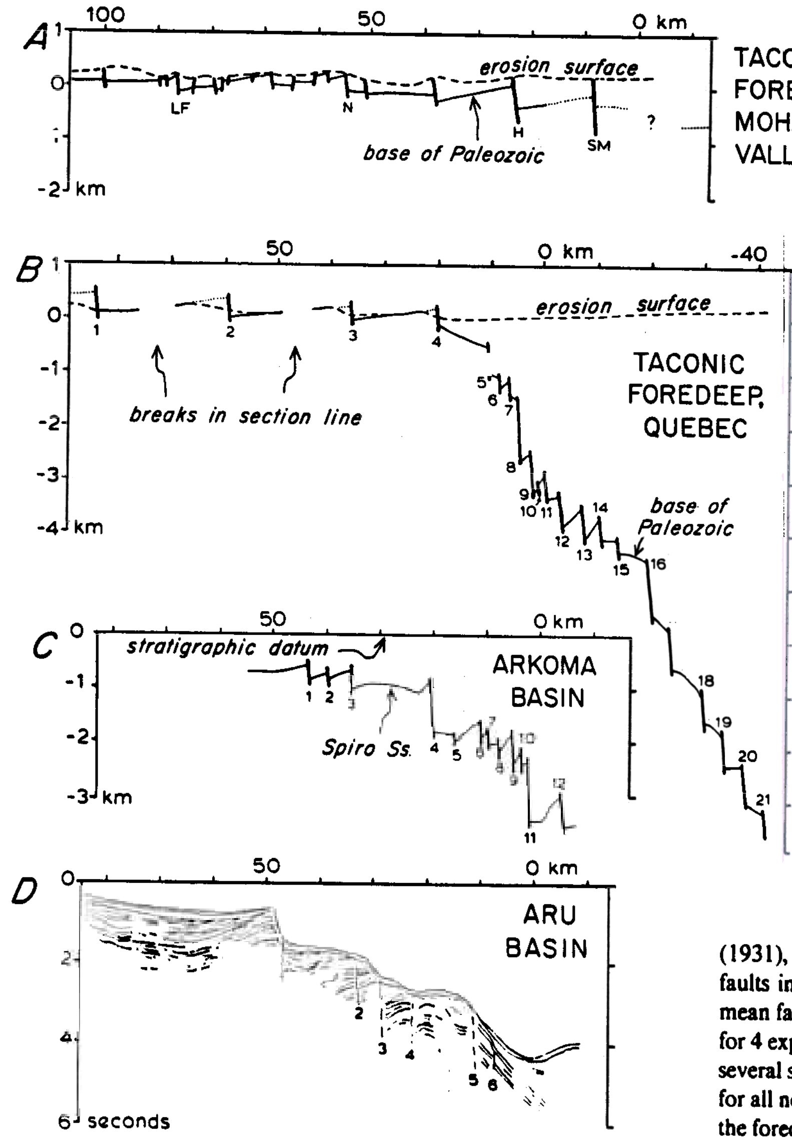

Figure 11. Comparative profiles across normal faulted foredeeps. The profiles are drawn at the same vertically exaggerated scales and are aligned horizontally with 0 km at the thrust front. (A) Taconic foredeep, Mohawk Valley, New York, based on this study. Datum is present-day horizontal, and the offset marker (thin line, dotted where approximate) is the Proterozoic-Paleozoic contact; it is cut by faults shown as thicker lines. Dashed line is the present erosion surface (B) Taconic foredeep, Quebec, including the cross section in Figure 10a, plus four additional faults closer to the craton (see Fig. 2). (C) Arkoma Basin, Arkansas, adapted from Houseknecht (1986). Datum is a key bed in the Atoka Formation, and the offset marker is the basal Atokan Spiro Sandstone. (D) Aru Basin, Australia–Banda Arc collision zone (section A–A′ in Figure 12b, redrawn from a seismic reflection profile illustrated in Jacobson and others (1979), at approximately the same vertical scale (1.5 sec equals 1 km) as the other profiles. In each profile, note the stepwise descent of the underthrust plate beneath the orogenic belt.

tion is restricted to minor local displacement along the northern (but not the southern) part of the Saratoga-McGregor fault (Geragthy and Isachsen, 1981).

SUPPLEMENTARY OBSERVATIONS OF EXTENSION IN OTHER COLLISIONAL FOREDEEPS

Taconic Foredeep North of the Mohawk Valley

The Cambrian and Ordovician evolution of the entire Northern Appalachian sector of the former continental margin was broadly similar to that described for the Mohawk Valley: following subduction of an oceanic tract, an arc or arcs collided with the North American passive margin during the Taconic orogeny in Ordovician time (Fig. 5). Normal faults resembling those in the Mohawk Valley occur along the intermittently exposed Taconic foredeep as far as northern Newfoundland, about 2,000 km distant.

Fracture zones traceable as north-northeast–trending valleys across basement rocks of the eastern Adirondacks (Fisher and others, 1970; Isachsen and McKendree, 1977) show that the larger Mohawk Valley faults continue west of the southern Champlain Valley (Fig. 2). Hudson (1931), Swinnerton (1932), and Quinn (1933) studied equivalent normal faults in the western Champlain Valley. Quinn (1933, p. 123) reported mean fault dips of 60° to 70° for 7 localities, and a range from 48° to 65° for 4 exposed major faults; Quinn also calculated extensional strains across several sections. These observations contributed to our choice of a 60° dip for all normal faults in the Mohawk Valley cross sections. A profile across the foredeep at Chazy near the northern end of Lake Champlain crosses 7 normal faults in 42 km. Although most displacements are poorly constrained, all but one of the faults are synthetic, and stratigraphic throw on the largest is about 915 m. Both Swinnerton (1932, p. 414) and Quinn (1933, p. 120) noted gentle backtilting of strata between faults.

Two lines of evidence suggest that displacement occurred during Ordovician time, at least in part. Quinn (1933) noted that normal faults cut strata as young as the Caradocian Hortonville Shale (paleogeographically akin to the Utica Shale) to the west of the Taconic thrust front; however, the normal faults do not cut Taconic thrusts, which were emplaced during the late Caradocian. As in the Mohawk Valley, Ordovician carbonate conglomerates occur near some normal faults in the Champlain Valley. Mehrtens and Gleason (1988) suggested that these conglomerates (Lacolle Formation) represent submarine talus or debris-flow deposits shed from active fault scarps.

The Mohawk-Adirondack–Champlain Valley belt of normal faults terminates at the Ottawa-Bonnechere graben (Kay, 1942) (Fig. 2). This elongate basin strikes at a high angle to the orogenic front and has been interpreted as an aulocogen that originated during late Proterozoic rifting (Burke and Dewey, 1973, p. 420–421).

North of the aulacogen along the St. Lawrence River, orogen-parallel normal faults again dominate the structure of the cratonic flank of the foredeep (Fig. 2). The geology of the Taconic foredeep in Quebec has been summarized by St. Julien and Hubert (1975), Belt and Bussiers (1981), Hiscott and others (1986), and Mehrtens (1989b). Whereas the Taconic foredeep evolution is comparable to that of the Mohawk Valley, the subjacent passive-margin sequence barely appears at outcrop and attains significant thickness only under the thrust belt.

All of the major normal faults in the Taconic foredeep in Quebec downdrop toward the orogen. Exposures illustrated by Clark and Globensky (1973, 1975a, 1975b, 1976a, 1976b), Clark (1972), and Riva and Pickerill (1987) show that dips close to 60° are representative of faults in the exposed part of the section. The best known fault is the classic Montmorency fault (a continuation of the Neuville fault of Table 2) near Quebec City (Fig. 2). Normal "drag" associated with the fault resembles that in the Mohawk Valley; that is, downdropped strata dip as steeply as 45° away from the fault, whereas upthrown strata are subhorizontal (Riva and Pickerill, 1987). Normal faults also cut the autochthon beneath the thrust belt, as revealed by seismic reflection profiling constrained by a few drill holes (Fig. 10a).

The foredeep section shown in Figure 11B crosses 21 normal faults. This transect was pieced together from three segments, in order to stay within or near Paleozoic foreland deposits, where offsets could be measured. Normal faults on the section occur as far as 84 km toward the foreland from the thrust front, and as far as 42 km toward the hinterland, beneath the thrust belt. Stratigraphic throws average ~390 m, with a maximum of ~1,000 m.

As in the Mohawk Valley, the exposed normal faults in Quebec die out in Ordovician strata, suggesting that displacement occurred during Ordovician time. Detailed stratigraphic relations support this conclusion (Mehrtens, 1989b). Most compelling is the presence of coeval Trenton Group sections with different facies and thicknesses on either side of the Montmorency fault, only 30 m apart (Mehrtens, 1989b, p. 143–144 and her fig. 22). On the upthrown block, the Trenton Group is thinner (8 m preserved); whereas on the downthrown block, the Trenton is thicker (11.5 m). The upthrown Trenton also contains a slump horizon that is absent from the downthrown side. Mehrtens (1989b) illustrated regional and local isopach patterns in the Black River and Trenton Groups that record uneven subsidence of a fragmented carbonate ramp; the most likely interpretation is that normal faults caused this fragmentation. The interpretation of the SOQUIP seismic profile by St. Julien and others (1983) (Fig. 10a) suggests that on some faults beneath the thrust belt, prior movement occurred during deposition of the basal clastic rocks (inferred rift facies).

Arkoma Basin, South-Central United States

Like the Appalachian margin, the Ouachita passive margin of North America also formed by rifting in late Proterozoic time, but it escaped collision until later. The Arkoma Basin, the foredeep of the Ouachita orogen (Fig. 12a), developed in Carboniferous time during thrusting of an accretionary wedge (presumed to have been in front of a volcanic arc) onto the continental margin (Viele, 1979; Lillie and others, 1983; Houseknecht, 1986). Strike-parallel normal faults are prominent regional-scale structures along the cratonic flank of the Arkoma Basin in Arkansas and Oklahoma (Haley, 1976; Miser, 1954). Comparable normal faults in the subsurface near the thrust front have been recognized through exploratory drilling (Koinm and Dickey, 1967; Buchanan and Johnson, 1968; Houseknecht, 1986). The latter faults have larger displacements than the surface faults which are exposed farther north, and their well-documented displacements and histories provide important constraints on processes of flexural extension. Most normal faults in the Arkoma Basin step down toward the orogen; the few known antithetic faults are paired with synthetic faults, forming narrow grabens (Haley, 1976; Houseknecht, 1986, p. 332).

Figure 11C shows a section across the western Arkoma Basin, adapted from Houseknecht (1986); the section is based mainly on wireline logs of 41 wells. The section crosses twelve normal faults, the farthest being 42 km cratonward of the allochthonous thrust front; additional normal faults of relatively smaller displacement occur farther north (Fig. 12a). Along section A–A′, stratigraphic throws average about 510 m with a maximum of about 2,030 m. Along section C–C′, Buchanan and Johnson (1968) documented successive burial of normal faults by diachronous Atokan foredeep deposits (Fig. 10b). Houseknecht (1986, p. 335) showed comparable relations along section B–B′ (not illustrated here). Sections B–B′ and C–C′ reveal that the normal faults are synorogenic growth structures, that locus of faulting migrated cratonward with time, and that normal faulting predated thrusting at any given place. Pre-Atokan motion on the three normal faults shown in Figure 10b is unlikely, because strata below unit 5 do not thicken across faults. Houseknecht (1986, p. 336) stated that "northward migration of (normal) faulting was undoubtedly caused by northward advance of the thrust front," in exactly the manner postulated by Chadwick (1917) for normal faults in the Taconic foreland.

Normal faults also occur in the Black Warrior foreland basin, along strike from the Arkoma Basin, about 400 km to east of the area shown in Figure 12a (Weisenfluh and Ferm, 1984). These faults occur to the north of, and strike subparallel with, the grain of the Ouachita Orogen, which here lies entirely in the subsurface beneath the Gulf Coastal Plain. Coal exploration has revealed evidence for Carboniferous movement on the normal faults. Older horizons in the Black Warrior coal measures show the greatest stratigraphic throws (as much as 300 m; Weisenfluh and Ferm, 1984).

High-angle faults in the extreme north of the area of Figure 12a also were active during the Ouachita Orogeny, but probably are not the product of flexural extension. These faults comprise a conjugate set; they strike northwest and northeast, oblique to contractional structures in the orogen and normal faults in the Arkoma Basin. Structures associated with certain of these faults resemble flower structure (both positive and negative), suggesting strike-slip or oblique-slip displacement. In the Tri-State lead-zinc district, movement along the northeast-striking Miami fault system during Chesterian deposition is evident from the presence of an anomalous thickness strata of that age within a downdropped block along the fault (Brockie and others, 1968, p. 412). This episode of displacement is significantly older (by about 20–25 m.y.) than the normal faulting in the Arkoma Basin, but it is coeval with the beginnings of Ouachita orogenesis because it occurred at a time when the slope-rise region of the North American passive margin was being inundated by flysch (Stanley Shale and Jackfork Formation; Houseknecht, 1986), presumably because the margin had neared or entered the approaching trench. During Chesterian time, the Tri-State district probably was located more than 200 km from the convergent plate boundary. In the present-day Australia–Banda Arc collision zone, the comparable position would be along the axis of, or cratonward of, the Sahul Rise (Fig. 12b; see below). In summary, although the oblique faults appear to be a foreland effect of Ouachita orogenesis, they are much older than the normal faults which likely formed by flexural extension, and occur significantly cratonward of the area that later was extended.

Timor Trough and Aru Basin, Australia-Banda Arc Collision Zone

Prominent syncollisional normal faults cut the outer slope of Timor Trough and its easterly continuation, Aru Basin (Fig. 12b), which together comprise the active foredeep of the Australia–Banda Arc collision zone (Von der Borch, 1979; Jacobson and others, 1979; Karig and others, 1987). Tertiary convergence between the Australian plate and Banda Arc first consumed a wide ocean and eventually led to arc–passive-margin collision beginning in the Pliocene. At least 150 km of Australian continental crust has been subducted (Karig and others, 1987; Veevers and others, 1978).

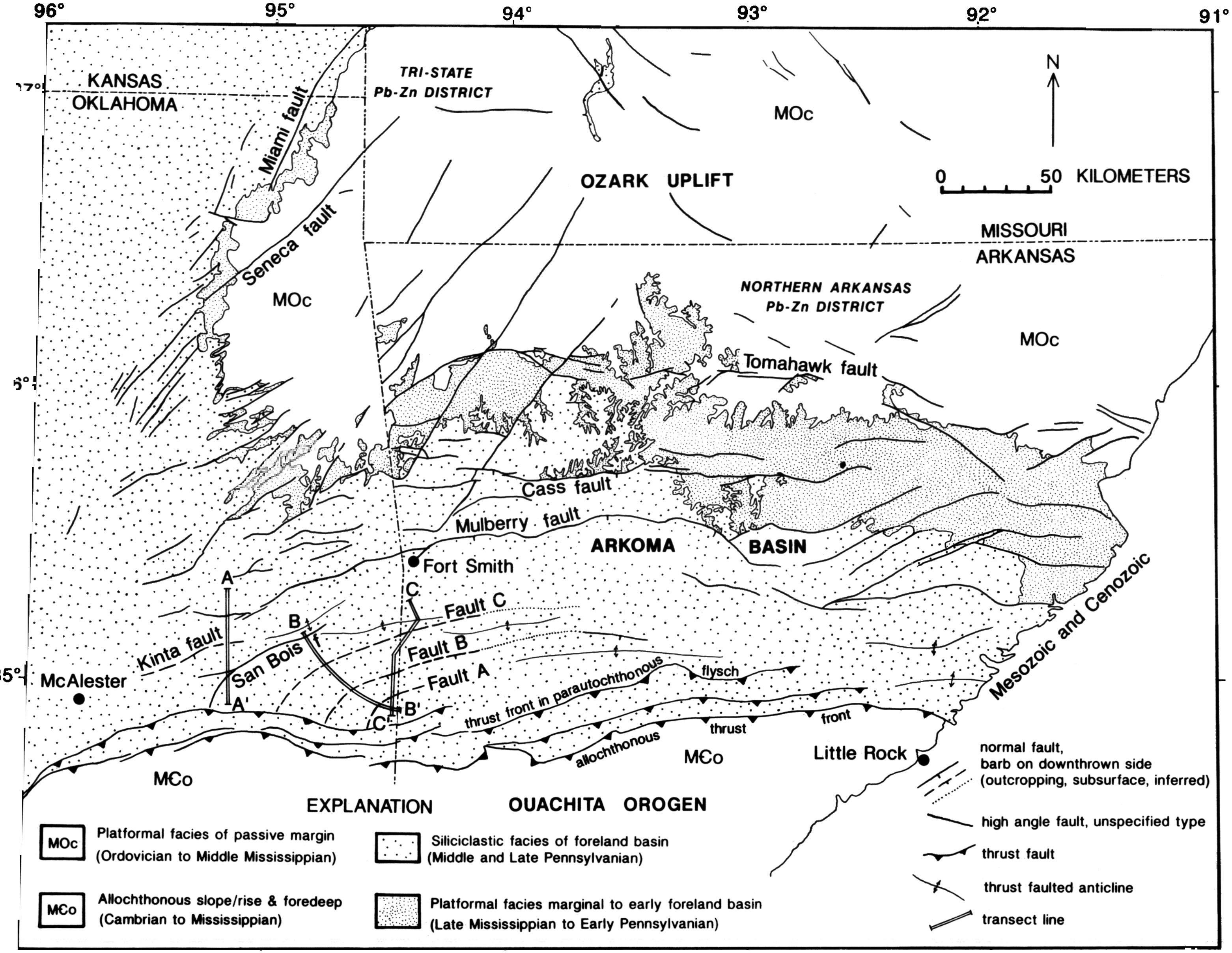

Figure 12a. Generalized geologic map of the Arkoma Basin. East-striking normal faults of interest occur throughout the Arkoma Basin; the largest normal faults occur in the subsurface just north of the thrust front, within a belt of faulted anticlines. Northeast- and northwest-striking high-angle faults to the north of the Arkoma Basin are older than the east-striking normal faults of Atokan age. Compiled from Haley (1976), Miser (1954), Anderson (1979), Jewett (1964), and Houseknecht (1986).

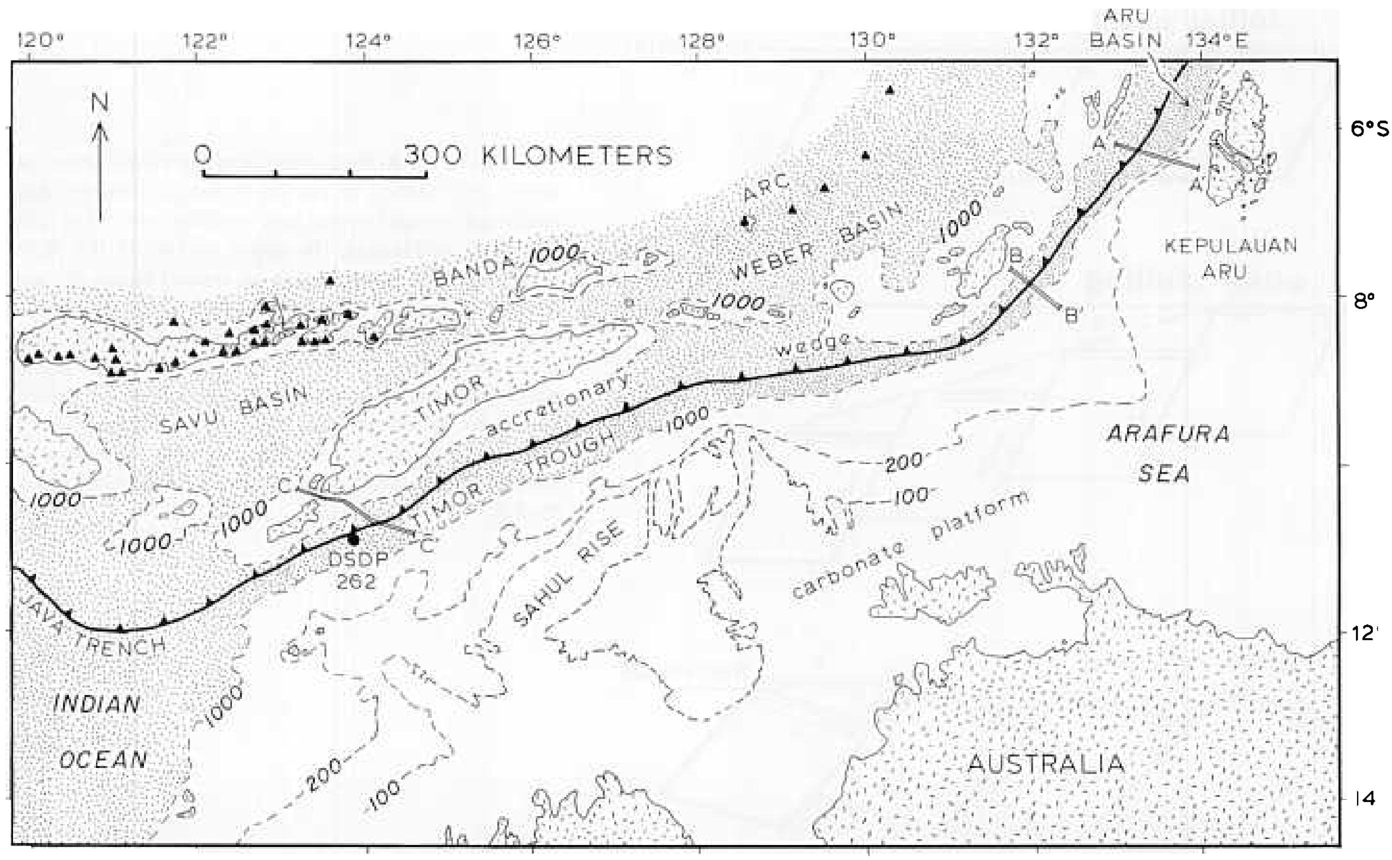

Figure 12b. Map of the Australia–Banda Arc collision zone, and the associated collisional foredeep, Timor Trough–Aru Basin (from Hamilton, 1979). Normal faults related to flexure of the Australian plate occur along most of the southern flank of Timor Trough. Sahul Rise and Kepulauan Aru mark the discontinuous forebulge. Random dashes, land areas; stipple, water depths >1,000 m; triangles, volcanoes of the Banda Arc. Isobaths (in meters) are dashed.

Normal faults have been imaged on most, but not all, of the many seismic reflection profiles across the active foredeep. Figure 11D reproduces an interpretation of such a profile from Jacobson and others (1979) through the Aru Basin, between the thrust belt and what appears to be an emergent forebulge, at Kepulauan Aru (Fig. 12b). A seismic reflection profile along C–C′ (reported by Karig and others, 1987, p. 19) is qualitatively similar; both profiles show a predominance of faults that downdrop toward the thrust belt. Seismic reflectors along profile C–C′ suggest that some of the active normal faults may have nucleated on Mesozoic passive-margin rifting faults (Karig and others, 1987).

Other Collisional Foredeeps

Ordovician-age normal faults also occur in the Taconic foreland in Newfoundland (Knight and James, 1987); parallel faults in the adjacent Gulf of St. Lawrence (Sanford and Grant, 1990) are probably of the same age. The influence of these faults on carbonate-platform sedimentation during passage of the Taconic forebulge was described by Knight and others (1991). Syncollisional, orogen-parallel normal faults also occur in the Carboniferous Variscan foredeep of Ireland (Mitchell, 1985), the Cretaceous foredeep of Oman (Harris and Frost, 1984), the Tertiary Venezuelan Basin (Harding and Tuminas, 1989), the late Paleozoic Uralian foredeep (Nalivkin, 1976), and in the Silurian Ellesmerian foredeep of Arctic Canada and North Greenland (Trettin, 1979; Surlyk and Hurst, 1984). The economic significance of such structures bears noting: Pb-Zn mineralization is localized along normal faults in Newfoundland (Lane, 1984, 1989) and Ireland (Mitchell, 1985), whereas normal faults in the Arkoma Basin, Oman, and Venezuela are associated with hydrocarbon fields.

GENERALIZATIONS ON FLEXURAL EXTENSION

Stratal Rotations and Normal Fault Geometry

Despite some differences, the Taconic foredeep, Arkoma Basin, and Aru Basin profiles share an important feature: most or all of the faults downdrop toward the orogen. Only in the Mohawk Valley is there a significant population of antithetic faults, and these have mostly smaller displacements. In each of these foredeeps, the underthrust plate descends stepwise beneath the fold-thrust prism. This is a very different near-surface configuration from that shown in Figure 1c, which is characterized by nearly symmetrical horsts and grabens produced by planar, downward-

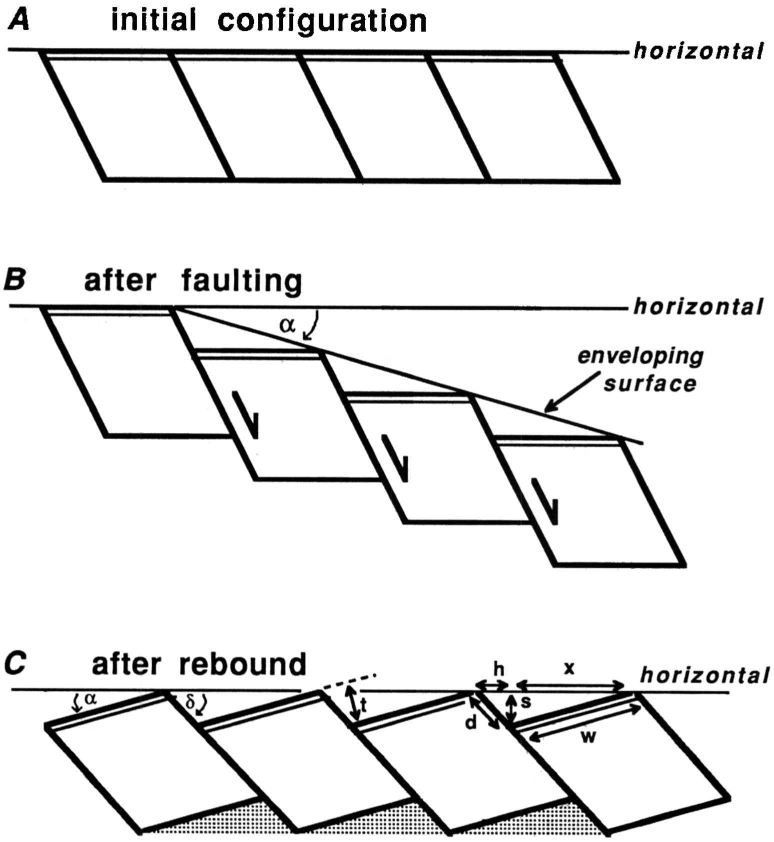

Figure 13. Relationships between simple shear, pure shear, and rotation as applied to flexure-induced planar-rotational normal faulting, and variables referred to in text. (A) Prior to faulting, the upper surface of the plate is horizontal. (B) After motion on several faults, the upper enveloping surface of the plate dips α to the right, but the tops of individual fault blocks remain horizontal. Shaded areas are where gaps would form as a result of faulting, which presumably would be accommodated by deformation at or below the base of the fault blocks. (C) After rebound, the upper enveloping surface is restored to horizontal, but the tops of individual fault blocks dip α to the left.

terminating normal faults. We suggest, instead, that the surface geometry of flexurally induced normal faults can be approximated by a row of toppled dominoes. This conclusion follows from the near-surface cross-sectional geometry alone and applies whether the faults dip 40° or 80°, and whether the displacements are 50 or 500 m.

In rift settings, tilt-blocks can be produced by either listric or planar-rotational faulting, but not by planar-irrotational faulting (see Wernicke and Burchfiel, 1982, for nomenclature). Whether listric or planar-rotational, faulting results in rotation of bedding with respect to horizontal. In foredeeps, it is necessary to consider the complication that an inclined rather than horizontal plate is extended. The meaning of "rotational normal faulting" is blurred unless the reference frame is specified. Two reference frames bear on fault geometry in cross section: (1) horizontal and (2), more important to the present paper, the enveloping surface of the flexed plate itself.

Figures 13A–13C show a geometric model which is compatible with the surface geology and which illustrates possible relationships between block rotation and normal faulting in a hypothetical foredeep. Dip-slip displacement on planar normal faults results in the stepwise configuration in Figure 13B. Faulting is irrotational with respect to horizontal, but rotational with respect to the upper enveloping surface. In Figure 13B, the dip of the enveloping surface, α, is given by $\alpha = \tan^{-1}(d/w)$, where d is displacement and w is initial fault spacing. Rebound of the enveloping surface to horizontal imparts bedding dips of α in the opposite direction (Fig. 13C). The net effect of Figures 13A–13C is exactly that illustrated by Jackson (1987, p. 8) in decomposing horizontal extension on planar-rotational normal faults (implicitly in a rift setting) into components of simple shear followed by bulk rotation. Figure 13B would represent an active foredeep with a bathymetric deep, such as Timor Trough, whereas Figure 13C would represent an ancient foredeep after rebound, such as the Mohawk Valley. Alternatively, Figure 13 could have been drawn with listric rather than planar faults. Listric normal faulting of an inclined slab would be accompanied by rotation with respect to both horizontal and the enveloping surface. Unfortunately, because fault-induced stratal rotations are so small in the examples studied, we are unable to determine whether normal faulting is planar-rotational or listric, only that it is one or the other and not planar-irrotational.

Stratal backtilting on the upthrown side of the leading major normal fault is an interesting feature common to three of the profiles (Fig. 11). This phenomenon cannot be the simple consequence of motion on faults that flatten at depth, because the footwall of the leading fault should not be involved in rotation in either the horizontal or enveloping surface reference frames. Strata in the footwall of fault 1 in the Aru Basin (Fig. 11D) dip gently toward the foreland. At a comparable position in the Arkoma Basin, footwall strata next to fault 1 dip gently toward the foreland, with respect to a late-orogenic key bed in the Atoka Formation (Fig. 11C). In the Mohawk Valley, the footwall of the main Little Falls fault (that is, the last major normal fault toward the foreland) is backtilted about 2° near the fault, declining to zero about 10 km to the west (see Figs. 3 and 6 and sections in Cushing, 1905). Such backtilting may be a manifestation of isostatic rebound of the tectonically unloaded footwall, as discussed by Buck (1988) and Wernicke and Axen (1988). In the Mohawk Valley, footwall uplift also is indicated by conglomerates within the Black River and Trenton Groups, which record real uplift of the clast sources with respect to sea level.

Magnitude of Flexural Extension

Extension in normal fault terranes results from the interplay of fault displacement and rotation. For the foredeeps of interest, we used the following approximate solution: $h = s/\tan\delta$, where h is heave (horizontal component of displacement) on a given fault, δ is present fault dip, and s is the difference in elevation across the fault between matching subhorizontal horizons. Using s as an approximation for throw, t, ignores any minor rotational effect of faulting, but the resulting error is much smaller than the probable error due to inaccurate fault dips. Application of this equation to each fault on the Mohawk Valley, Quebec, and Arkoma Basin sections yielded the values in Tables 1 through 3. We assumed fault dips of 60° for all sections based on outcrop evidence in the Mohawk and Champlain Valleys and Quebec, but some faults could be 5° to 10° steeper. On the other hand, there undoubtedly are many undetected minor faults, especially along the Quebec and Arkoma Basin sections, which are based mainly on subsurface data. These sources of error will tend to offset each other.

Figure 9 relates properties of the normal faults to distance across strike on the three sections. Large-displacement faults tend to occur close to and beneath the orogen (Fig. 9a). This trend presumably results from the convergence-driven migration of the zone of normal faulting, such that the frontal normal fault at any given time was the youngest, and faults closer to the orogen were active longer.

Cumulative heave (Σh) increases toward the orogen on each section (Fig. 9b). Σh is greatest for Quebec (about 4,375 m), smaller for the Arkoma Basin (about 2,950 m), and smallest for the Mohawk Valley (about 1,200 m). Most of the difference is due to the fact that the sections cross different parts of their respective foredeeps. For example, the Mohawk Valley section includes no faults beneath the foredeep axis or thrust belt, a zone on the other two sections with the largest normal faults. Overall percentage extension also tends to increase toward the orogen, with some variations reflecting the sampling of different parts of the foredeeps. Extension across the Mohawk Valley profile is small (0.6%), and only somewhat larger (1.25%) for the portion east of the Little Falls fault. In the Champlain Valley, close to the orogen, the longest section has about 4.5% extension. In Quebec, the overall extension is 3.5%, but beneath the orogen (between faults 5 and 21), it is about 7.25%.

TABLE 3. NORMAL FAULTS IN THE ARKOMA BASIN, OKLAHOMA, TRANSECT A-A′

Fault*	Distance to thrust front (km)†	Throw (m)§	Polarity	Cumulative heave (Σh) (m)**
(1) Mulberry	42	221	Synthetic	185
(2) Mulberry	39	85	Synthetic	234
(3) Mulberry	35	302	Synthetic	409
(4) Kinta	20	949	Synthetic	957
(5)	15	106	Synthetic	1,018
(6)	11	289	Synthetic	1,185
(7)	9	234	Synthetic	1,320
(8)	7	255	Synthetic	1,467
(9)	5	638	Synthetic	1,835
(10)	3	255	Synthetic	1,983
(11) St. Bois	2	1,106	Synthetic	2,621
(12)	−4	575	Synthetic	2,953

*Numbers refer to faults in Figure 11c.
†Measured to thrust front in parautochthonous flysch.
§Data for estimates of displacement from Houseknecht (1986, his Fig. 6).
**Heave calculated assuming all faults have 60° dips.

Magnitude of Vertical Displacement Associated with Flexure

A plot of cumulative stratigraphic throw versus position (Fig. 9c) shows that normal faults in the Mohawk Valley, Quebec, and Arkoma Basin alone could have produced water as deep as in Timor Trough. The simple geometric model in Figure 13b implies that such a correlation should exist. This correlation suggests that the normal faulted upper part of the lithosphere accommodated itself to the top of an unfaulted elastic lithosphere. Plots such as Figure 9c provide a potential way to study paleobathymetry in ancient foredeeps that are cut by normal faults, a long-standing problem in paleogeographic analysis (Shanmugam and Walker, 1980).

Normal faults facilitated subsidence in these foredeeps because of an overwhelming predominance of synthetic structures (Fig. 9c). Only one antithetic fault occurs in Quebec (off the section), a few small ones occur in the Champlain Valley sections transitional to the Mohawk Valley area, and a few occur in the Arkoma Basin (also off the sections). Antithetic faults are common only in the Mohawk Valley, where, of 27 faults crossed by the section, nine dip cratonward. Were antithetic and synthetic faults equally abundant, the curves in Figure 9c would plot as jagged subhorizontal lines hugging the x-axis. Instead, because down-to-basin faults predominate, cumulative down-to-trench displacement increases toward the orogen in all three examples.

Nature and Depth of Lower Termini of Normal Faults

The rotation (with respect to the enveloping surface) of normal fault blocks at the surface implies either (1) flattening at depth of faults into one or more low-angle detachment(s) or zone(s) of distributed simple shear or (2) termination of faults at the base of the lithosphere. As discussed in more detail under "Implications for Mechanics of Lithospheric Flexure," the second possibility cannot account for co-occurrence of normal faults and forebulges in trenches and foredeeps, and is therefore improbable. From available surface data, we cannot clarify the details of the first possibility, but for purposes of illustration in Figure 6D, we show the Mohawk Valley normal faults as planar structures that terminate in a zone of ductile simple shear below the brittle upper crust ("detachment," below). Conservation of cross-sectional area is achieved in Figure 6D by offsets at the base of the brittle fault blocks equivalent in magnitude to offsets at the surface. Voids thereby created presumably could be filled through some combination of ductile and brittle deformation (possible solutions to the room problem were reviewed by Axen, 1988). Alternatively, the normal faults might have been drawn as listric structures. Geologic field relations and seismic reflection profiles in the Basin and Range have been interpreted by some workers (for example, Davis and Lister, 1988; Allmendinger and others, 1983) to show shallow-level normal faults (which bound fault blocks with domino-style surface configuration) soling into low-angle detachments at depth. Although the surface dips of tilted blocks and the magnitude of extension are very different, the toppled domino surface geometry of the Taconic, Arkoma, and Timor-Aru foredeeps is qualitatively similar to that in parts of the Basin and Range. Our extension estimate suggests that beneath the thrust front on the Mohawk Valley section, the amount of net translation between the base of the normal faulted layer and subjacent rocks is only about 1 km; furthermore, this value declines to zero at the leading normal fault (Fig. 6D). Thus, any postulated detachment at depth in the Mohawk Valley, or in the other normal faulted foredeeps studied, must have displacements which are many times smaller than the major extensional detachments in the Basin and Range (Wernicke, 1985).

The Mohawk Valley normal faults are shown as ending at about 17.5-km depth, on the basis of indirect evidence from analogues. In oceanic subduction zones, bending-induced normal fault earthquakes occur at depths as great as 25 km, whereas the few relevant deeper earthquakes are compressional (known examples are ~45, 40, and 19 km below the sea floor; Chapple and Forsyth, 1979). Chapple and Forsyth (1979) concluded that in bending oceanic lithosphere, a neutral surface exists somewhere between about 25 and 40 km. Comparable earthquake data from flexed continental lithosphere have not been published but, by analogy, we suggest that a neutral surface also exists at some depth in such settings. A likely possibility is that normal faults in foredeeps continue at least to the depth at which ductile deformation predominates over brittle. We estimate this transition to lie at depths of 15 to 20 km in the Mohawk Valley (the intermediate value of 17.5 km is shown in Fig. 6D), based on the follow-

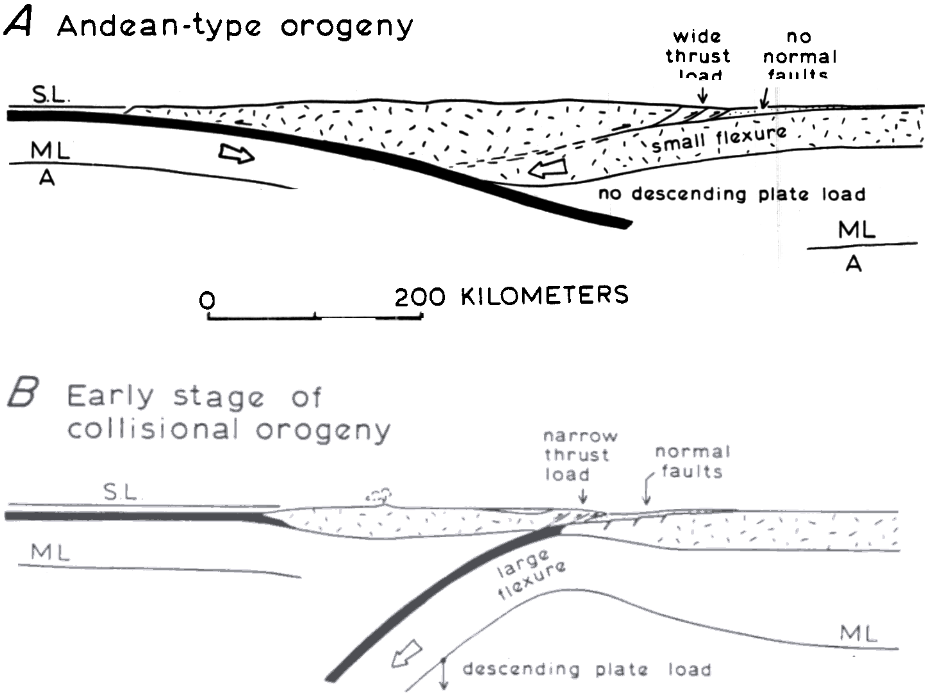

Figure 14. Generalized cross sections comparing Andean-type foreland basins, where the angle of flexure in the outer slope of the foreland basin is too gentle to cause normal faulting, and arc-continent collision settings, where the flexure is larger and normal faulting occurs.

ing assumptions: (1) Grenville quartzo-feldspathic gneisses occur at depth and probably deform ductilely at temperatures above 300 °C (Carter and Tsenn, 1987), and (2) the present geothermal gradient in the Mohawk Valley region (20 to 15 °C/km; Putman and Young, 1985, p. 61) corresponds to a depth to quartz ductility of 15 to 20 km. The present gradient probably is not significantly different from that during normal faulting, because any minor thermal effect of late Proterozoic rifting would have decayed by the Caradocian, some 100 m.y. later (Vitorello and Pollack, 1980), and because there is no evidence for significant thermal disturbances in the past 150 m.y. (M. Heizler, 1989, personal commun.).

A prominent zone of gently dipping to horizontal seismic reflectors is present at a depth range of about 15 to 20 km on a COCORP profile across the eastern Adirondacks (Brown and others, 1983, line 7). These reflectors continue to the east for some distance under the thrusts of the Taconic allochthon (Brown and others, 1983, line 3) but cannot be traced west beyond the projected extension of the last significant normal fault in the Mohawk Valley (Brown and others, 1983, lines 7 and 8). We suggest that some reflectors in this zone might correspond to the shear zone which we infer lies at depth in the Mohawk Valley.

Comparison with Other Convergent Systems and Significance of Slab Dip

Normal faults have been observed in the outer slopes of most deep sea trenches, including the Japan, Bonin, Tonga, Philippine, Mariana, Peru-Chile, and Aleutian trenches (Jones and others, 1978). Most faults have down-to-trench displacements (Jones and others, 1978), which tend to increase toward the trench and only rarely exceed 1 km (Bradley and Kusky, 1986). The similarities with normal faults in foredeeps are striking.

Not all oceanic subduction zones and collisional foredeeps are cut by normal faults, however. The Nankai Trough is one well-documented deep-sea trench where faults are absent in the outer trench slope; Karig and others (1987) attributed their absence to a relatively lower angle of flexure of the oceanic plate. Normal faults are absent in Taiwan Straits, the thickly sedimented active foredeep of the Taiwan-China collision zone, and at some sections through Timor Trough (for example, line B–B′ in Fig. 12B) (Jacobson and others, 1979). Jacobson and others (1979, p. 216) suggested that the Australian passive margin has ruptured wherever it has been bent through an angle of more than about 1.5°. A somewhat larger critical angle (exceeding the 2.7° flexure at the Nankai Trough) seems to be required for rupture of oceanic lithosphere.

Flexure-induced normal faults are lacking or poorly developed in Andean-type foreland basins; for example, normal faults are not reported in the foreland basins of the Canadian Rockies or central Andes. Flexure of the foreland of the Canadian Rockies has imposed a regional dip of 0.4° to 1.4° in strata beyond the thrust front (sections in Ollerenshaw, 1978). A similar range of slopes occurs in the equivalent part of the central Andean foreland (Smith, 1989; Roeder, 1988). These observations suggest that the continental lithosphere in Andean-type forelands is bent through a gentler angle than in collisional forelands. Following a suggestion by Karig and others (1976) for oceanic subduction zones, we suspect that this is due at least in part to the size (width) of the thrust load, wide loads leading to proportionally smaller deflections than narrow ones. During arc-continent collision, furthermore, a loaded passive margin follows an oceanic slab down to considerable depths, and the downgoing plate passes through a bend of 30° or more, compared with much smaller total bends in Andean settings (Fig. 14).

DISCUSSION

Recognition of Flexural Extension in Regional Tectonic Analysis

In the Taconic foredeep from New York to Newfoundland, the Arkoma and Black Warrior Basins, and Timor Trough-Aru Basin, orogen-

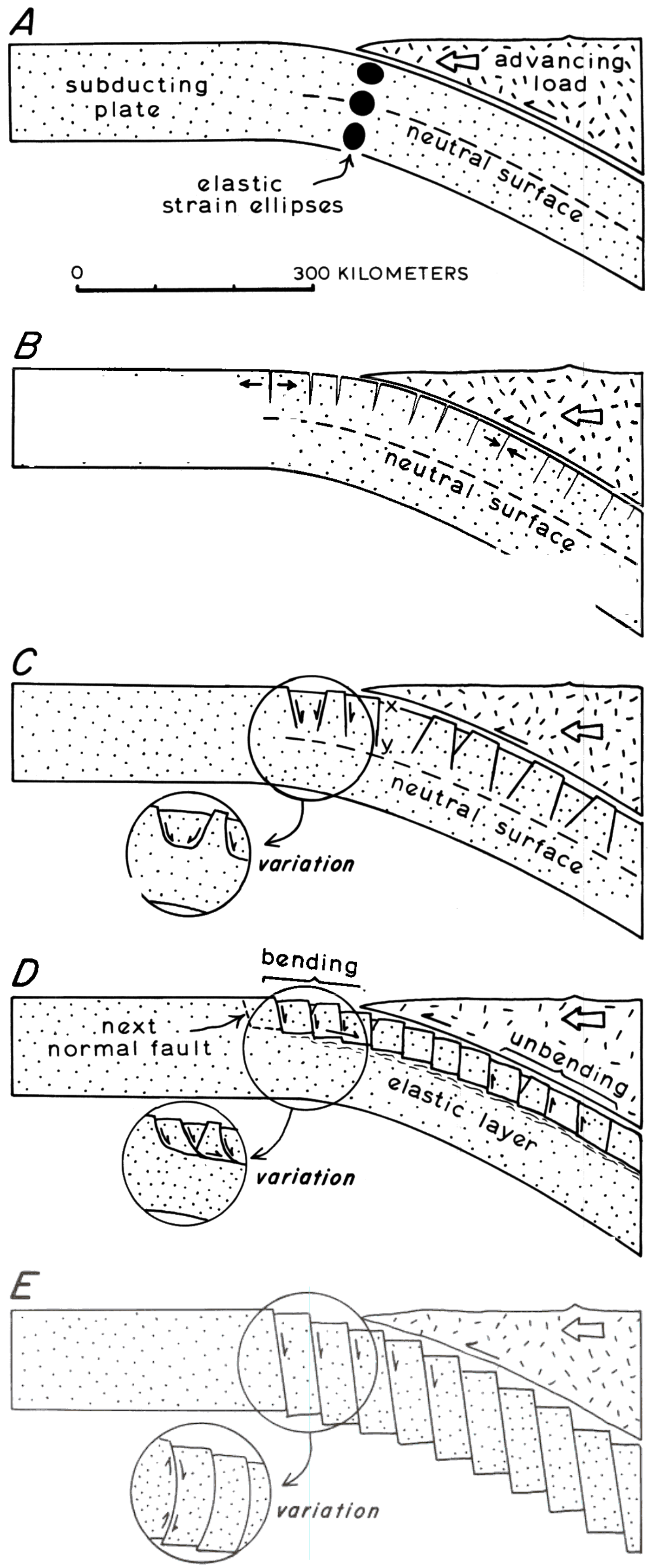

Figure 15. Five alternative models of lithospheric flexure in collisional foredeeps. (A) Lithosphere modeled as a vertically and horizontally isotropic, elastic beam. Elastic strain, shown schematically by ellipses, is entirely recoverable. (B) Extension in the convex outer part of the flexed lithosphere is accommodated by opening of vertical, downward-terminating dilatant cracks. (C) Extension is accommodated by motion on downward-terminating normal faults. This creates space problems at depth because corresponding displacements must also die out (the offset at the surface at point x must die out to zero at point y). Inset shows a variation with synthetic and antithetic faults soling into a subhorizontal detachment where shear-sense confrontations must occur. (D) Extension is accommodated by motion on planar-rotational normal faults, which sole into a detachment or ductile shear zone of unspecified thickness, at mid-crustal depths. Inset shows a variation with listric faults. (E) Straight or curved normal faults penetrate the entire lithosphere. In all models, plate convergence results in concomitant migration of the extensional front. For clarity, none of the alternative models shows the possible effects of contractional deformation on lower-plate rocks beneath the orogenic load.

parallel normal faults are the dominant structures over vast areas. These faults occur along passive margins that formed by rifting, and therefore, the age of faulting is an important question in regional tectonic analysis. Normal faults in collisional foredeeps might conceivably have moved during prior rifting, during collision, after collision, or some combination. Stratigraphic evidence favoring the second possibility includes facies and thickness changes across faults in syncollisional strata, and overlap by flysch or molasse. A consequence of plate convergence is that the normal faulting front moves across the downgoing plate, in advance of the plate boundary. We visualize this as a steady-state configuration. In ancient arc-continent collision zones, normal faults across a foredeep transect are preserved at four major stages of development: (1) the first incipient faults of a few meters displacement, located 80 or more km cratonward of the thrust front; (2) major growth faults on the outer slope; (3) normal faults buried beneath flysch along the foredeep axis; and finally (4) normal faults tectonically buried beneath the fold-thrust belt. The orogen-parallel strike of flexural normal faults allows them to be discriminated from another class of synorogenic normal faults, which strike perpendicular to the orogenic front, and are explained by Şengör and others' (1978) impactogen model (see also Hancock and Bevan, 1987).

On a lithospheric scale, extension induced by flexure differs fundamentally from that related to rifting, and the corresponding structural expressions also are distinctive. Rifting (whether by "pure shear" or "simple shear"; Wernicke, 1985) involves extension of the entire lithosphere. Regional extension measured in tens of kilometers is typical of rift settings, and fault-block surfaces are typically rotated many tens of degrees. Rifting typically is associated with magmatism and often leads to formation of a passive continental margin. In contrast, flexural extension involves only the upper lithosphere, below which there is probably a regime of contractional deformation. Flexural extension is characterized by regional extension of a few kilometers, and by stratal backtilting (with respect to the enveloping surface) of a few degrees at most. At least in the Phanerozoic, associated magmatism is absent or minor (but see Hoffman, 1987, for several enigmatic proposed examples of Proterozoic foredeep magmatism). It is notable that none of the models for lithospheric flexure in Figure 15 provides any obvious mechanism for the generation of melts; this problem needs careful study.

Implications for Mechanics of Lithospheric Flexure

The present study adds structural geologic constraints to modeling studies of lithospheric flexure, which heretofore have been based primarily on indirect observations (bathymetry, gravity, and seismicity). Flexure-induced normal faults occur in both oceanic and continental lithosphere, and regardless of composition of the flexed plate, there appears to be no qualitative difference in the brittle response of the lithosphere, whether oceanic or continental, to large-amplitude deflections.

Five alternative models for the structural geology of lithospheric flexure are shown in Figure 15. Figure 15A depicts the lithosphere as a vertically and horizontally isotropic, elastic beam, following the approximation used by Walcott (1970) and numerous later workers in mathematical flexure models. Despite the success of the elastic model in simulating many first-order observations, it violates a basic fact: normal faulting in trenches and foredeeps shows that the upper part of flexed lithosphere can be deformed beyond its elastic limit, as Chapple and Forsyth (1979, p. 6734) noted (see also Glasner and Schubert, 1985). Nearly half a billion years after the bending moment was applied in the Taconic foredeep, and despite rebound of the foredeep axis almost to sea level, fault offsets of hundreds of meters still exist. Hence lithospheric-scale deformation was permanent and, by definition, *inelastic*, at least in the upper crust. The assumption of elasticity could be reconciled with permanent deformation by special pleading (for example, if normal faulting is confined to a relatively thin upper portion of the flexed plate), but more complex multilayer models with vertically anisotropic, quasi-elastic lithosphere (for example, Chapple and Forsyth, 1979; Bodine and others, 1981) are likely to be more realistic approximations.

Figure 15B improves on Figure 15A by allowing flexural extension to cause brittle failure, but it is inconsistent with field observations. In Figure 15B, tension cracks terminate downward, toward a neutral surface. That this is kinematically possible is proven by the experimentally deformed slab in Figure 1a, but the process clearly cannot be extrapolated from the laboratory to lithospheric plates, because the primary evidence for flexure-induced extension is the occurrence of normal faults; gaping tension cracks are not observed in foredeeps.

Figure 15C is a minor modification that solves the tension-crack problem but creates others. One shortcoming of the model in Figure 15C is that the symmetrical horst-and-graben fault geometry is inconsistent with the observed step-wise descent of flexed lithosphere beneath the overriding plate. Synthetic normal faults in trenches and foredeeps are more numerous and have larger displacements than antithetic faults. The model in Figure 15C also implies that large normal fault displacements, up to 1 km in a number of cases, simply die out down dip. For example, displacement at the surface at point X would have to diminish to zero at point Y, with unacceptable loss of cross-sectional area. One might argue that the normal faults are listric rather than planar (Fig. 15C inset), soling into one or more subhorizontal ductile shear zones at depth. This would lead to shear-sense confrontations along the shear zone(s), however. On the other hand, something must happen to the normal faults at depth; otherwise they would penetrate into the concave side of the neutral surface implicit in the model.

Figure 15D shows a new model that avoids all of these problems. In this model, step-wise normal faults in the downgoing plate are rotational in the reference frame of the enveloping surface, but irrotational with respect to horizontal (compare Fig. 13B). This model accounts for the predominance of synthetic faults and the correlation between cumulative throw in the Taconic and Arkoma foredeeps with water depths in Timor Trough. The faults might terminate in a number of ways at depth; we show two end-member variants, following Brun and Choukroune (1983, p. 353). One possibility is that the faults are planar and terminate at a subhorizontal zone of ductile deformation where any space problems (the gaps between toppled dominoes) are accommodated. Another possibility is that the faults are listric and merge with a detachment (Fig. 15D, inset). Regardless of details at depth, the rotational configuration of the fault blocks requires net displacement between the upper-crustal fault blocks and subjacent rocks, the fault blocks moving relatively toward the orogen. The magnitude of displacement along the detachment would necessarily increase toward the orogen at each normal-fault intersection.

Figure 15E shows an alternative geometry wherein planar normal faults penetrate the entire lithosphere. This model can account for the stepwise descent of the faulted lower plate, and in fact differs only qualitatively from Figure 15D in the depth at which normal faulting is accommodated. One problem with Figure 15E is that it cannot account for compressional earthquakes at depth (Chapple and Forsyth, 1979), but curved faults resembling escalator stairs (Fig. 15E, inset) could account for both the earthquakes and the stepwise descent of the plate. A glaring problem with both variations, however, is that they imply a virtual decoupling of the loaded from the unloaded part of the underthrust plate, across numerous steep faults of negligible strength. If this were the case, lithospheric flexure could not be approximated by the bending of an elastic beam under an applied load, and the occurrence of peripheral bulges seaward of virtually all trenches would become very difficult to explain, inconsistent with the strong correlation documented by Caldwell and others (1976). It is possible that lithosphere-penetrating normal faults cut the subducted plate beneath the forearc, where most Benioff zones bend downward, but we doubt that this model is applicable to the outer trench slope of subduction and collision zones.

SUMMARY

Foredeeps formed by arc–passive-margin collision are zones of extension, which is induced by bending of the underthrusting plate beyond its elastic limit. In the Taconic and Arkoma foredeeps, normal faults dominate the structure of zones paralleling the thrust front and extending 80 km or more toward the craton. A predominance of down-to-basin (synthetic) faults leads to a stepwise descent of the underthrust plate beneath the fold-thrust belt. At least in the Taconic and Arkoma foredeeps, normal faults alone appear to have been entirely capable of producing water as deep as in Timor Trough. This suggests to us that the normal faulted upper part of the lithosphere in these places accommodated itself to the top of an unfaulted elastic lithosphere.

In contrast with many published depictions, we note that normal faults cut flexed lithosphere into a series of fault blocks that are gently rotated with respect to the upper enveloping surface of the faulted plate. We suggest, further, that these faults may be linked at depth by a subhorizontal detachment or zone of ductile simple shear. Extension is small compared to that in rift settings; the upper crust of a 100-km-wide foredeep adjacent to a thrust belt might typically be stretched by only a few kilometers as a result of bending.

In its response to large-amplitude flexure, continental lithosphere in collisional foredeeps does not appear to differ qualitatively from oceanic lithosphere in trenches. Therefore, marine seismic reflection profiles across normal faulted trenches contain a wealth of additional information bearing on flexural extension, worthy of detailed structural analysis.

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